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The Main Peculiarities of the Environmental Distribution of Microfossils in the Riphean Basins of Siberia

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Abstract – Previously, the study of the paleoenvironmental distribution of Precambrian microfossils was mainly based on actual interpretations of the biotas composed of silicified forms, which used to exist in extremely specific shallow zones of paleobasins. In this paper we considered these problems by using direct data on the paleoenvironments of the fossiliferous sequences and analyzing the paleoecological settings of the Riphean microfossils (organic-walled and silicified ones). This approach provided new means of assessing the facial and ecologic distribution of ancient microorganisms within a wide range of paleoenvironments in the Riphean basins of central (the Turukhansk and the Uchur-Maya regions) and northern Siberia (the Anabar massif). All the regions revealed some uniform, spatially regulated pattern (structure) of the microfossils' lateral distribution; four distinctive paleoenvironmental groupings of ancient forms have been discriminated: the deep-water, "optimal (medium deep)," transitional, and shallow-water ones. We demonstrate the microphytological specific features of these groupings in the Riphean basins of siliciclastic, clay-carbonate, and carbonate sedimentation. The taxonomic composition, specific ecological and evolutionary traits of these microfossil groupings, and their stratigraphic potential were examined. The water depth was regarded as the main factor responsible for the distribution of microfossils in these paleobasins. The richest and most diverse were "optimal" groupings of microfossils in the medium-deep zones of the Riphean basins dominated by large, complicated, chiefly eucariotic forms. The obtained data do not correspond to previous notions concerning the high rates of evolutionary transformations of Riphean eucaryotic communities predominantly concentrated in the medium-deep zones. We suggest that the revealed pattern of lateral distribution of ancient microorganisms had a long-term history of formation that terminated in the Riphean.

INTRODUCTION

For a long time, Precambrian microfossils were studied with the main aim of solving stratigraphical problems on the basis of empirically derived peculiarities of taxa's (to be more exact, parataxa's) time-distribution. The impact of paleoecological factors on the space and time distribution of biotas or communities was not taken into account. Such an approach helped us to understand the biostratigraphical potential of microfossils (for review see Yankauskas *et al.*, 1989; Semikhatov *et al.*, 1990); however, today there is every reason to contend that the potentials of this approach have been exhausted. It seems much more promising now to analyze the integrated features of paleoenvironments and microbiotic populations to reconstruct paleoecosystems and their specific organization.

There have been various attempts to understand the main peculiarities of Precambrian microfossil paleoecological distribution. But, so far, such attempts mainly, or exceptionally, dealt with silicified microfossils (e.g., Hofmann, 1976; Knoll *et al.*, 1991), which, for the most part, represent specific impoverished communities of benthic mat-forming microorganisms confined to extremely shallow zones of carbonate platforms. As a result, such paleoecological analysis, in fact, neglected certain, much more diversified microfossil communities from siliciclastic sequences, where microfossils are represented by organic-walled forms. Nevertheless, Knoll (e.g., 1982, 1984) proposed a general model of lateral and ecological distribution of microfossils, based predominantly on the actualism principle in interpretation of silicified microorganisms. According to that model, the impoverished, chiefly benthic groups of procaryots occupied, as a rule, only the peritidal zone and were replaced towards the open-marine setting zones by more diverse communities dominated by plankton eucaryotic forms. It was the latter communities that revealed distinct evolutionary transformations contrasting with the extreme morphological conservatism of silicified microfossils from shallow environments. Considering that conclusion, Knoll attributed the general trend of increasing dimensions of the Proterozoic coccoidal and filamentous microfossils (Schopf, 1977) not to the evolution of ancient microorganisms, but to global changes in the distribution of basin landscapes (to the relatively increased area of the open-marine settings in the Late Riphean).

Recent detailed studies of the ecological structure of the Riphean and Vendian organic-walled microfossils from several regions of North Eurasia (e.g., Veis and Semikhatov, 1989; Burzin, 1993; Vidal and Nystuen, 1990) immediately demonstrated the necessity to verify Knoll's model. Continuation of these studies on the

Riphean sections in Siberia (Veis and Vorob'ieva, 1992; Veis and Petrov, in press) revealed some definite spatially regulated pattern in the lateral distribution of organic-walled microfossils. Using original sedimentological and microphytological investigations, we provide a more detailed substantiation of the above phenomenon. Our data are based on the most favorable parts of the Riphean successions of the Turukhansk uplift, the Anabar massif, and the Uchur-Maya region (Fig. 1) that contain diverse organic-walled and silicified microfossils and demonstrate distinct facies variations. In addition, the Turukhansk and Anabar sections were thoroughly examined by one of the authors sedimentologically (Petrov, 1993a, 1993b; and unpublished data). The position of the relevant formations in the composite sections of the above regions is presented in Fig. 1.

When reconstructing the paleoecological structure of the fossil communities, we primarily used the results of our sedimentological analysis of the sequences containing such communities, and the confinement of particular biotas to settings revealed by our analysis. The actualism model in ecological interpretation of individual members of such biotas was regarded to be of secondary importance for the analysis of their environmental settings. The previously published data on the composition of the organic-walled microfossils encountered in the described deposits (Timofeev *et al.*, 1976; Veis, 1988; Pyatiletov, 1988; Veis and Vorob'ieva, 1992; and others) were supplemented by new materials from our studies. These materials proved to be the essential to the biotas from the Ust'-Il'ya and Kotuikan formations of the Anabar massif. Data on the Turukhansk uplift and the Uchur-Maya region were expanded only in quantity. To make the picture more

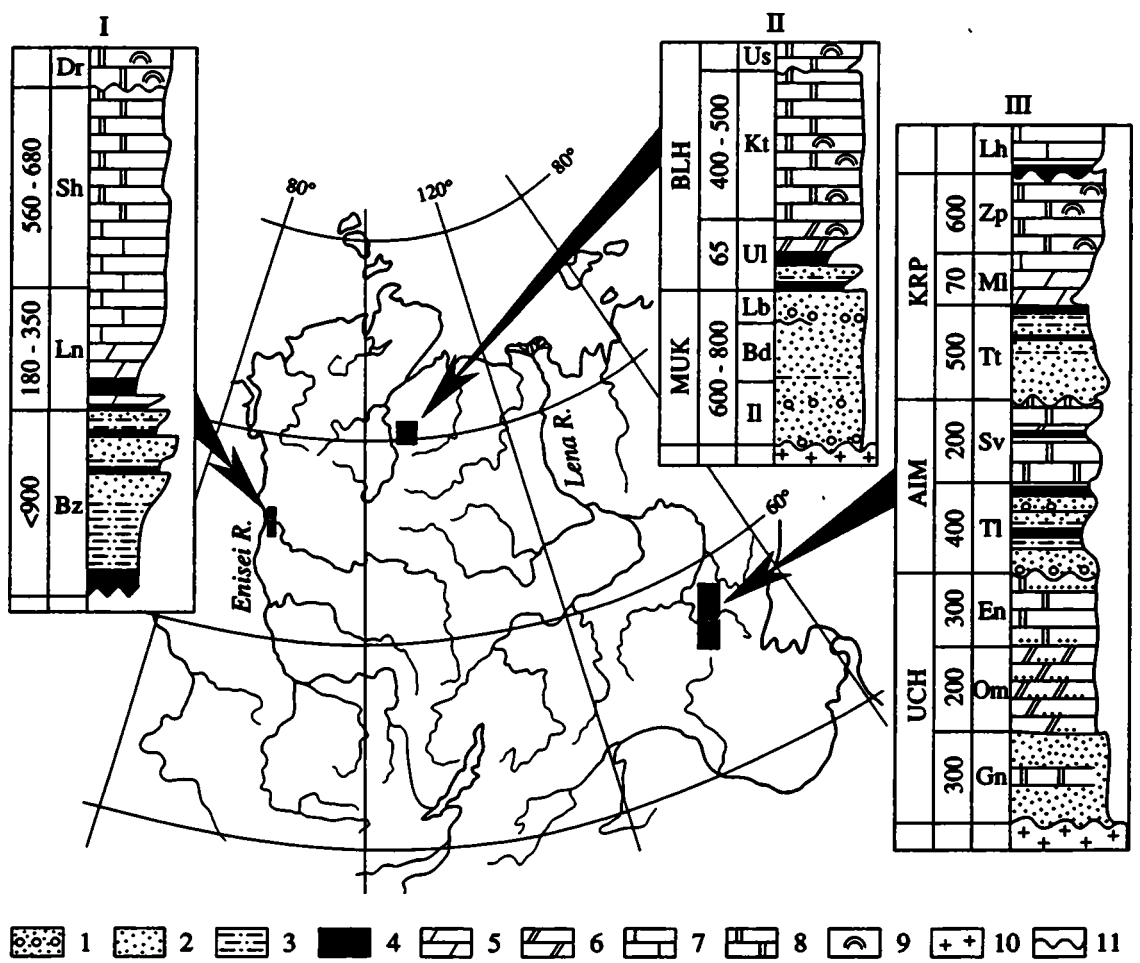

Fig. 1. Location of examined Riphean sections and stratigraphical position of the units under consideration.

1 - 8 – prevailing composition of rocks: 1 – gravelstones, grits, and coarse-grained sandstones; 2 – coarse-grained sandstones; 3 – sandstones and siltstones; 4 – siltstones and shales; 5 – clayey limestones; 6 – clayey dolomites; 7 – limestones; 8 – dolomites; 9 – prevailing of stromatolites; 10 – crystalline basement; 11 – regional disconformities. Groups: MUK – Mukun; BLH – Billyakh; UCH – Uchur; AIM – Aimchan; KRP – Kerpyl'. Formations: Bz – Bezymyannyi; Ln – Linok; Sh – Sukhaya Tunguska; Dr – Derevnya; Il – Il'ya; Br – Burdur; Lb – Labaztakh; Ul – Ust'-Il'ya; Kt – Kotuikan; Us – Yusmastakh; Gn – Gonam; Om – Omnya; En – Ennin; Tl – Talyn; Sv – Svetlyi; Tt – Totta; Ml – Malgina; Zp – Tsipanda; Lh – Lakhanda. Regions: I – Turukhansk; II – Anabar; III – Uchur-Maya. The thickness of stratigraphical units is given in meters.

complete, in addition to organic-walled microfossils we also considered silicified forms reported from the carbonate units under consideration. Their composition is described according to published data (Golovenok and Belova, 1981, 1984, 1993; Sergeev, 1993) and some unpublished materials on the Sukhaya Tunguska Formation of the Turukhansk uplift, which were kindly given to us by V. Sergeev.

GEOLOGICAL POSITION AND AGE OF THE EXAMINED DEPOSITS

The most representative data were obtained for the Turukhansk uplift and the Anabar massif. Below, these data are discussed in more detail than those on the Uchur-Maya region.

The Turukhansk uplift, situated along the western margin of the Siberian Platform, displays a succession of Riphean deposits that have a rather significant time-range, and are unconformably overlain by the Vendian and Cambrian strata. Our paper will deal with the lower part of this section, including the siliciclastic Bezymyannyi Formation, the clay-limestone Linok Formation, and the carbonate Sukhaya Tunguska Formation. These formations constitute a transgressive sequence, bounded at the top by a regional erosional surface, which preceded the accumulation of the Derevnya Formation. The deposits are exposed within two submeridionally elongated, closely spaced tectonic sheets and form gently westward-dipping monoclines. The described material was collected from the outcrops of the above formations along the Nadporozhnaya, Kamennaya, Nizhnyaya Tunguska, and Sukhaya Tunguska rivers.

The Anabar massif is a typical intraplatform gentle uplift. Its Riphean deposits are unconformably overlain by Vendian and Cambrian beds and rest subhorizontally. Among the deposits exposed, here we shall describe the uppermost horizons of the siliciclastic Mukun Group, the carbonate-clay Ust'-Il'ya Formation, and the essentially carbonate Kotuikan Formation in their best studied section along the Kotuikan River.

As for the Uchur-Maya region, we shall analyze only two formations from its thick and continuous Riphean sequence, namely the terrigenous-carbonate Omakhta and the siliciclastic Totta formations (Fig. 1). The Omakhta Formation corresponds to middle horizons of the Uchur Group, the basal one in the Riphean section of the region. The Totta Formation is the basal unit in the Kerpyl Group. In the composite section of the region, it is separated from the Uchur Group by the terrigenous-carbonate Aimchan Group and unconformably overlaps all older rocks down to the Archean basement. Below, we shall consider only several sections of the Omakhta and the Totta formations that are most important to the context of this paper, and which are exposed within the limits of the gentle platform depressions (the Uchur and Maya depressions respectively) in the southwestern part of the region (Semikhatov and Serebryakov, 1983). The description of the Omakhta Formation is based on its sections along the Idyum River, the southern limb of the Uchur depression; data on the Totta Formation were obtained from a slightly wider area along the Aim, Omnya, and Maya rivers located in the eastern limb of the Maya depression. In terms of their sedimentology, these deposits were investigated, however, less thoroughly than those from the Turukhansk uplift and the Anabar massif. Still being formed in a wide range of environments, they are rich in microfossils and therefore useful for comparative analysis.

The geological age of most of the formations has been treated differently in different works. This may be explained not so much by the lack of objective age information, but rather by the range of views on the criteria for tracing the Middle–Upper Riphean boundary in Siberia. Now, addressing the relevant information, it is important to emphasize that the succession of the Bezymyannyi, Linok, and Sukhaya Tunguska formations in the Turukhansk uplift is firmly correlated with the succession of the three formations forming the Kerpyl Group of the Uchur-Maya region, due to striking similarities in their superposition and paleontological characteristics (Fig. 1). By the same criteria, the Derevnya Formation is correlated with the lower part of the Lakhanda Group resting on the Kerpyl Group. This unambiguous correlation expands the data base relevant to the age evaluation of the above deposits.

The isotope-geochronological data on the Turukhansk region is very scarce. Some K-Ar datings on glauconite obtained in the 1960s from different parts of this section vary within the limits of 970 - 800 Ma and seem to testify only to reorganisation of the respective isotope systems about 800 - 850 Ma ago. The information on the isotopic age of the deposits under consideration is slightly broader for the Uchur-Maya sections. As for the Totta Formation, there are obviously rejuvenated K-Ar and Rb-Sr datings of glauconite, indicating the time of regressive katagenesis. Furthermore, there is a series of K-Ar datings of the same mineral equal to 1170 - 970 Ma (Semikhatov and Serebryakov, 1983). When combined with the available K-Ar datings for the Aimchan and Lakhanda deposits, these data strongly suggest that the time-range of the Kerpyl Group (and hence, of the correlative Bezymyannyi, Linok and Sukhaya Tunguska formations) falls approximately within the limits of 1200 - 1000 Ma (Semikhatov and Serebryakov, 1983; Khomentovskii *et al.*, 1985; Komar, 1990; Shenfil', 1991). This conclusion and some representative microbiotas from the Derevnya and Lower Lakhanda deposits, which are similar to those from the base of the Upper Riphean type section (the Muldakaevo biota), along with characteristic Upper Riphean taxa of stromatolites, led several scientists to conclude that the Middle–Upper Riphean boundary in the Siberian sections should be placed between the Kerpyl and Lakhanda groups and their analogs (see Semikhatov and Serebryakov, 1983; Semikhatov, 1991; for review and bibliography). However, some geologists place this boundary lower in the section, at the base of the Kerpyl Group and its analogs

(Komar, 1973, 1990; Khomentovskii *et al.*, 1985; Veis, 1988; Shenfil', 1991). Their judgement is based on: the appearance of certain dimensional and morphological varieties of organic-walled microfossils in the Bezymyannyi and Totta deposits, several stromatolite taxa known from the Upper Riphean in the South Urals and occurring in the Upper Kerpyl and Linok–Sukhaya Tunguska sequences, and/or historical-geological considerations. It should be stressed that this very position of the Middle-Upper Riphean boundary in Siberia corresponds to that in the South Urals, if the boundary is considered microphytologically.

Until recently, the age of the Ust'-Il'ya and Kotuikan formations was unanimously considered to be the Early Riphean. This conclusion was based on the composition of the Kotuikan stromatolites and the K-Ar datings of glauconite from the Mukun, Ust'-Il'ya, and post-Kotuikan (Yusmastakh) deposits (Komar, 1966; Zlobin and Golovanov, 1970; Semikhatov and Serebryakov, 1983). This inference was supported by recently obtained Rb-Sr and K-Ar isochronous datings of Al-glauconite from various horizons of the Ust'-Il'ya deposits. They respectively correspond to 1491 ± 5 and 1465 ± 10 Ma and are considered representative of the time of early diagenesis of these sediments (Gorokhov *et al.*, 1991). According to Veis and Vorob'ieva (1992), this conclusion is disputed by new data on the composition of the Ust'-Il'ya and Kotuikan organic-walled microfossils, among which there are several dimensional and morphological types considered to be characteristic of the Upper Riphean deposits in other Siberian sections (the Totta Formation and younger units). At the same time, Pokrovskii and Vinogradov (1991) regarded initial $^{87}Sr/^{86}Sr$ ratios in the Kotuikan and Yusmastakh carbonates as more consistent with their position in the upper horizons of the Early Proterozoic. We also have different opinions on the age of the Ust'-Il'ya-Kotuikan and Bezymyannyi-Totta deposits.

As for the Omakhta Formation of the Uchur-Maya region, all geologists regard it, together with the entire Uchur Group, as the Early Riphean in age. The grounds for this conclusion are characteristic stromatolites, some K-Ar datings, and the stratigraphic position of the formation beneath the Middle Riphean Aimchan Group. The Omakhta assemblage of organic-walled microfossils does not contradict this opinion, though some taxa recognized here were previously known only from younger deposits (Veis and Semikhatov, 1989).

DEPOSITIONAL ENVIRONMENTS AND DISTRIBUTION OF MICROFOSSILS

The Turukhansk Uplift

As was mentioned above, the Riphean succession of the region corresponds to a large transgressive cycle, which includes three conformably occurring formations with total visible thickness of up to 1800 m (Fig. 2).

The Bezymyannyi Formation is a unit of frequently alternating siltstones, shales, and sandstones with a visible thickness of up to 900 m. The unit is divided into three subformations. The lower of these is dominated by dark grey shales and siltstones with thin interbeds of fine-grained sandstone. The middle subformation consists of rhythmically alternating shales, siltstones, and fine- to coarse-grained sandstones. The upper subformation mainly includes light-grey coarse-grained sandstones. The section structure and sedimentation environments of the Bezymyannyi deposits have been described previously (Petrov, 1993b). Here we will review only features that are important to the context of this paper.

According to the available data, the siliciclastic Bezymyannyi sediments accumulated within an open sea basin deepening to the northwest. The accumulation proceeded chiefly below or near the activity base of long-period storm waves. The general sedimentation pattern was determined by the high-energy flows generated by storms.

Against a background of generally increasing amounts of relatively coarse-grained sandstones in the up-section, the Bezymyannyi Formation displays three large shallowing cycles (Fig. 2), complicated by small-scale (10 - 20 m) regressive cycles. This hierarchy of cycles reflects the unidirectional evolution of the paleobasin, the lateral migration of its facies zones, and the pulsating siliciclastic input. Against the background of these events, the floor relief of the Bezymyannyi paleobasin gradually flattened, owing to the frequent accretion of the siliciclastic shelf, as can be concluded from the analysis of the contemporaneous section intervals throughout the entire distribution area of the Bezymyannyi Formation.

The large-scale cycles seem to be the result of a significant (over a hundred kilometers) lateral migration of facies zones of the paleobasin. The sections of such cyclic units demonstrate some directional changes in the depositional environments and therefore, in terms of paleoecology, they are the most informative elements of the formation (in our reconstructions, the depth of the paleobasin is regarded as the main environmental factor). As to the small-scale cycles, they reflected certain short-term and low amplitude oscillations of the sea level (probably never exceeding 10 m) that did not alter the general facies arrangement of the Bezymyannyi basin. The short-term oscillations promoted only rapid and frequent injections of sedimentary material to the off-shore parts of the basin, where they were fading with depth increase. In this connection, the small-scale cycles can be regarded as elementary acts of the directed pulsational accretion of the Bezymyannyi shelf.

The Linok Formation conformably overlies the Bezymyannyi Formation. Being 180 - 380 m thick, it is divided into three subformations (Fig. 2). The lower subformation (10 - 50 m) is represented by greenish-grey clayey limestones, marls, calcareous siltstones, and shales prevailing in its lower part. The middle

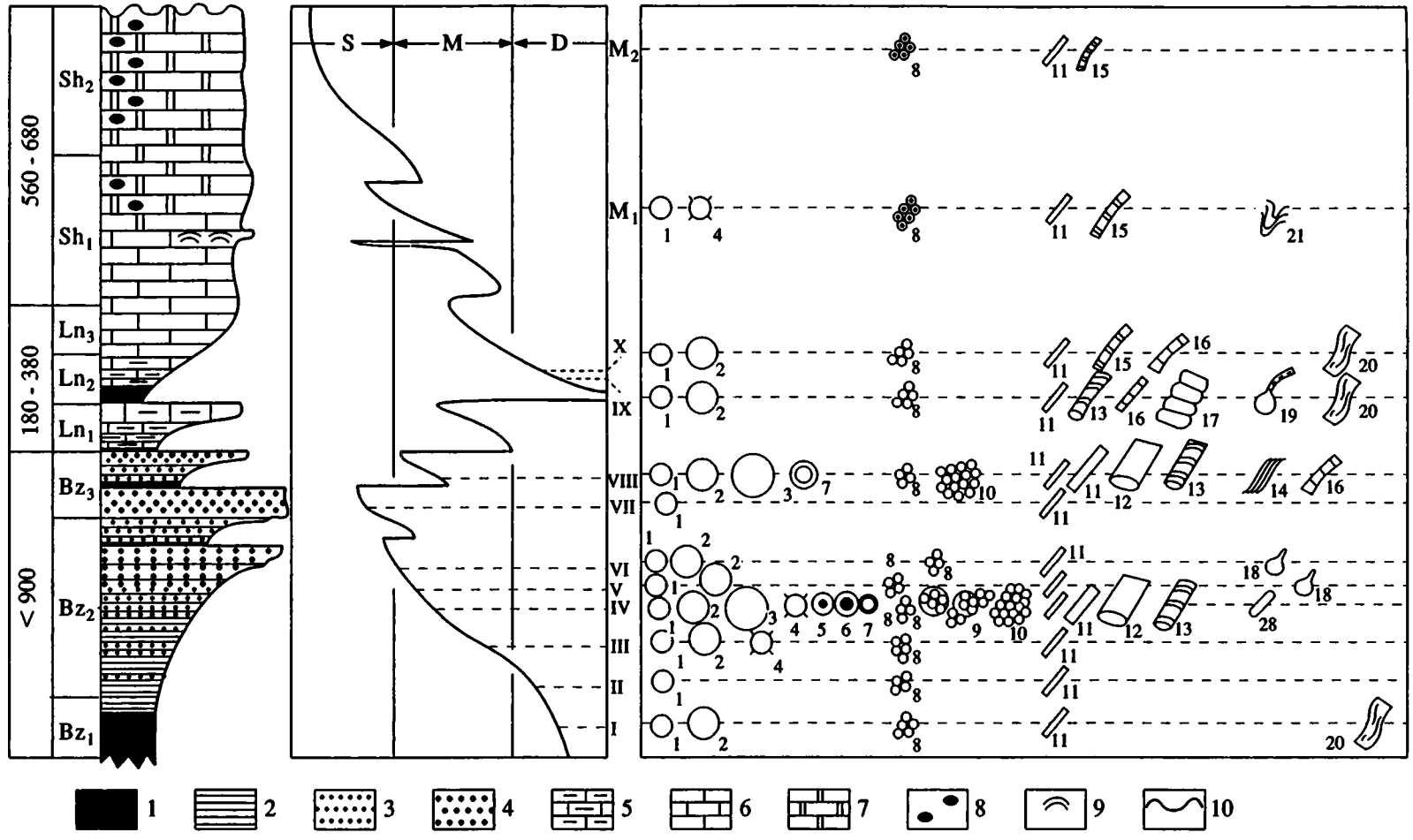

Fig. 2. Distribution of microfossils in the three lower formations of the Riphean succession of the Turukhansk uplift: in the Bezymyannyi (Bz), Linok (Ln), and Sukhaya Tunguska (Sh) formations and the change in depositional environments during their accumulation.

Paleosettings: S – shallow-water; M – medium deep; D – deep-water. The thickness of the formations is given in meters. 1 - 7 – rock composition; 1 – siltstones and shales; 2 – siltstones, shales, and fine-grained sandstones; 3 – fine- and medium-grained sandstones; 4 – medium- and coarse-grained sandstones; 5 – clayey limestones; 6 – limestones; 7 – dolomites; 8 – chert nodules; 9 – stromatolite buildups; 10 – erosional surface. Microfossils (figures in the picture) – see the explanation to Fig. 3. I - X – occurrence levels of organic-walled microfossils; M_1 and M_2 – occurrence levels of mineralized microfossils.

subformation (40 - 120 m) consists of black shales and their dark-grey calcareous and silico-calcareous varieties. The upper subformation (140 - 220 m) is dominated by clayey limestones, with their clay content decreasing upward in the section.

According to the analysis of the structure, the Linok succession reflects two large transgressions, each of which was followed by a long-term stage of shallowing well expressed in the rock sequence structure (Fig. 2). The first transgression took place at the beginning of the Linok time, which was the turning point in the history of the Turukhansk paleobasins, as it indicates the transition from siliciclastic to carbonate sedimentation. The second transgression, preceding the accumulation of the middle subformation, did not change the general type of sedimentation, but stipulated the appearance of the deepest water environments known for the time interval under study.

When examined, the structural successions of the Linok Formation section show the sediments of its lower and middle subformations to be formed in a relatively deep-water setting, below the wave base, while the upper subformation deposits reflect a gradual decrease in the paleodepth, when they were exposed to storm wave activity (Petrov, 1993a). Let us consider in more detail the sedimentation environments of the middle subformation, which yields nearly all the organic-walled microfossils known from the Linok Formation.

The deposits of this subformation were formed under calm hydrodynamic conditions below the storm waves base and, on the whole, are products of clay sedimentation inside the basin with periodic injections of siliciclastic silt-size material. This material was concentrated in thin (under 1 mm) continuous laminae of two types: the laminae, with occasional gradational bedding and occasional signs of erosion at their base, and thinner laminae composed by well-sorted grains of silt size. Most likely, the former resulted from suspension flows generated by strong storms, and the latter, from eolian-transported continental material. The many shales and carbonate deposits in the middle subformation are characterized by their fine and discontinuous black banding produced by thin (less than 1 mm) dark films oriented parallel to the layers, which occasionally yield intrinsic crinkled lamination. The films are enriched in organic matter and silt-size clastic particles, and correspond to accumulation episodes of organic and fine siliciclastic material. Most probably, these films are remnants of microbial mats because their inner microstructures reveal a stage-to-stage growth, when they were trapping and binding fine silt particles (the crinkled lamination, solitary siltstone grains enveloped by films).

The Sukhaya Tunguska Formation is 540 to 650 m thick, and consists of limestones and dolomites divisible into two subformations of almost equal thickness (Fig. 2). The lower subformation (250 - 310 m) is composed of dark limestones and replaced by dolomites in the northern part of the region. The upper subformation (280 - 350 m) is dominated by light-grey dolomites with numerous concretions of early diagenetic cherts containing silicified microfossils (Mendelson and Schopf, 1982; Golovenok and Belova, 1992a, 1992b; 1993).

The structural analysis of the Sukhaya Tunguska carbonates shows that, on the whole, their sequence reflects a general shallowing with two distinct regressive cycles, which approximately correspond to two subformations (Fig. 2). Each cycle reveals a transition from relatively deep-water environments in settings below the wave base to shallow-water conditions with

Plate I. Microfossils from the Linok and Bezymyannyi Upper Precambrian formations of the Turukhansk uplift.

All specimens presented here are stored at the Geological Institute of the Russian Academy of Sciences, Moscow.

1 - 4, 6 - 30 – the Bezymyannyi Formation; 22, 26 – the lower subformation, the first level of microfossil occurrence (the upper reaches of the Kamennaya River, the left tributary of the Bol'shaya Shorikha River); 1, 3, 4, 6 - 11, 13 - 15, 19 - 21, 25, 27, 29 – the middle subformation; 14 - the third level of microfossil occurrence; 1, 3, 4, 6 - 10, 13, 21, 25 – the fourth level of microfossil occurrence (the left bank of the Nizhnyaya Tunguska River near the Voronikhin Mys); 20, 27, 29 – the fourth level of microfossil occurrence (the upper reaches of the Kamennaya River, the left tributary of the Bol'shaya Shorikha River); 11, 15 – the fifth level of microfossil occurrence; 19 - the sixth level of microfossil occurrence (the right bank of the Sukhaya Tunguska River, downstream the mouth of the Bechevnik Creek); 2, 12, 16 - 18, 23, 24, 28, 30 – the upper subformation, the eighth level of microfossil occurrence; 5 – the Linok Formation, the middle subformation, the tenth level of microfossil occurrence (the left bank of the Nizhnyaya Tunguska River near the Voronikhin Mys); 5, 22, 26 – the deep-water grouping of Turukhansk organic-walled microfossils; 11, 14, 15, 19, 27 – the transitional grouping of Turukhansk organic-walled microfossils; 1 - 4, 6 - 10, 12, 13, 16 - 18, 20, 21, 23 - 25, 28 - 30 – the "optimal" grouping of Turukhansk organic-walled microfossils. Bar for 1, 3, 4 = 500 μm; for 2, 8, 15, 18, 26 - 28 = 240 μm; for 5, 13, 19, 21, 23, 24 = 100 μm; for 6, 14 = 1000 μm; for 7 = 700 μm; for 9 = 80 μm; for 10, 12, 17 - 22, 29 = 400 μm; for 11, 16, 20, 25, 30 = 150 μm; 1, 3, 4, 6 – *Chuaria circularis* Walcott emend. Vidal et Ford; 1 – specimen (sp.) no. 2078/4250; 3 – sp. no. 2078/127; 4 – sp. no. 2078/1251; 6 – sp. no. 2078/123; 2, 17 – *Leiosphaeridia jacutica* (Tim.): 2 – sp. no. 905/74 and 17 – sp. no. 905/710; 5 – *Siphonophycus* sp.; a large sheath with closed rounded end and transverse structure; sp. no. 2078/7910; 7, 12 – *Nucellosphaeridium nordium* (Tim.): 7 – sp. no. 2078/1252 and 12 – sp. no. 905/727; 8 – a fragment of a broad thallus with transverse-wrinkled surface; sp. no. 2078/4211; 9 – pleurocapsoidal aggregate of subspheric colonies of coccoidal cells; sp. no. 2078/4244; 10 – *Myxococcoides grandis* Horod. et Donalds; sp. no. 2078/1253; 11, 16 – fragments of coenobial aggregates of small vesicles *Leiosphaeridia*: 11 – sp. no. 903/22 and 16 – sp. no. 903/55; 13 – *Pterospermopsimorpha capsulata* Jank.; sp. no. 2078/1219; 14 – a large vesicle with rare low rounded-conic processes, ? *Trachyhystrichosphaera* sp.; sp. no. 903/16; 15 – *Leiosphaeridia* sp.; vesicles at the stage of division; sp. no. 903/2454; 18, 29 – fragments of structureless corroded thalluses: 18 – sp. no. 905/71 and 29 –sp. no. 903/1959; 19, 20 – germinated reproductive structures of algae or fungi: 19 – sp. no. 905/221 and 20 – sp. no. 903/2015; 21 –*Eosynechococcus* sp., no. 2078/4245; 22 – damaged fossil of a broad thallus with longitudinal structure; sp. no. 903/65; 23 – *Polytrichoides lineatus* Herm.; sp. no. 905/76; 24 – *Leiotrichoides typicus* Herm., sp. no. 905/656; 25, 26 – *Asperatofilum experatus* (Herm.): 25 – sp. no. 2078/1257 and 26 – sp. no. 903/658; 27 – a filamentous fossil composed of small black bodies-granules (? *Polysphaeroides* sp.); sp. no. 905/225; 28 – *Oscillatoriopsis* sp.; fragment of a broad trichome with short cylindrical cells; sp. no. 905/76; 30 – *Eomycetopsis robusta* Schopf emend. Knoll, Golubic; sp. no. 905/64.

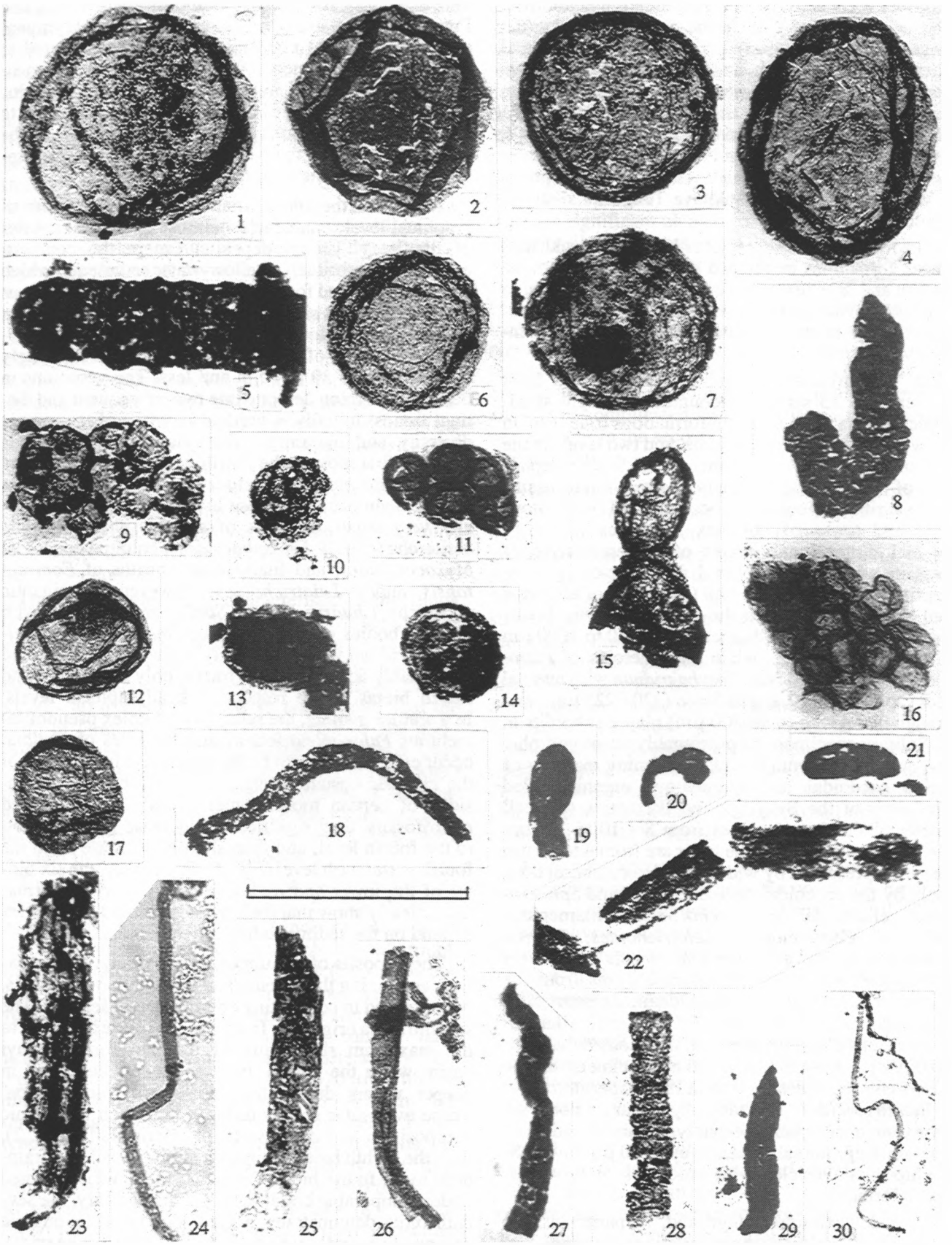
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Plate I.

permanent wave activity. The transition is evident from both the granulometry of clastic carbonate rocks (the change from fine- to coarse-grained wackestones), and the replacement of mechanogenic carbonate sedimentation by their biogenic *in situ* accumulation (stratiform stromatolites). Upward in the section, the role of the shallow-water sediments tends to increase. As a result, the upper subformation is dominated by shallow-water deposits of the carbonate shelf. It is these very deposits that yield the bulk of the Sukhaya Tunguska silicified microfossils.

Microfossil assemblages occur in all the Turukhansk Riphean formations mentioned above, though the most abundant and diverse are in the Bezymyannyi Formation (Veis, 1988; Veis and Petrov, 1994). Below we shall summarize the most important features of these assemblages vital to our arguments and conclusions.

The Bezymyannyi assemblage includes at least 20 genera and 25 species encountered at eight stratigraphical levels in all three subformations (one level in the lower, five levels in the middle, and two levels in the upper subformations). They include all basic morphotypes of Precambrian organic-walled microfossils, namely acritarchs, colonial coccoidal and filamentous forms, and those of complex structures (Plate I). The total dimensional ranges of representatives of these taxa are shown in Plate I. In terms of quantity, the acritarchs obviously prevail (over 50% of all forms found here). Among them the sphaeromorphic *Leiosphaeridia* and *Chuaria* (diameters from 10 to 1000 μm and more) predominate, while the spheroids of *Pterospermopsimorpha* and *Nucellosphaeridium* with internal bodies (up to 100 μm), edged *Simia* (220 - 225 μm), and sporangialike *Aimia* (over 400 μm) play a subordinate role. Solitary specimens of presumably acantomorphic forms are also present. The overwhelming majority of colonial coccoidal and filamentous organic-walled microfossils of the Bezymyannyi Formation (the cell diameter of the former varies from 8 - 10 to 30 μm, sometimes to 40 - 50 μm; the latter are from 4 to 70 μm wide) are represented by widespread Precambrian taxa, namely by the coccoidal *Myxococcoides* and *Sphaerophycus*; ellipsoidal *Eosynechococcus*; filamentous sheaths of *Eomycetopsis*, *Leiotrichoides*, *Polytrichoides*; and trichomes of *Oscillatoriopsis* (some over 40 μm wide). Complicated and rare forms from the Bezymyannyi Formation include: chainlike aggregates of coccoidal cells of *Polysphaeroides*; pseudofilamentous forms of *Palaeopleurocapsa*; fragmented thalluses of *Ostiana*; and such filamentous specimens as the crinkled-costate *Rectia*, ribbonlike *Asperatofolium*, and *Taenitrichoides*. The most significant among the Bezymyannyi complex morphotypes are broad thalluses with longitudinal striation (up to 90 μm in width) and ellipsoidal vesicles with processes of *Germinosphaera* forms.

The wide taxonimic range and, in particular, the dimensions of the most characteristic morphotypes of the Bezymyannyi organic-walled microfossils regularly vary within the stratigraphical levels mentioned above (Fig. 2). As has been demonstrated by Veis and Petrov (1994), these two parameters of the Bezymyannyi organic-walled microfossils are closely related to some paleoenvironmental factors revealed by sedimentological analysis of microfossil-containing deposits. The most significant of these was the depth, and, to a certain extent, also hydrodynamic intensity in the respective zones of the paleobasin. We will now exemplify these relations by actual data.

In general, the microbiotas of the above eight stratigraphical levels characterize almost all facies varieties of the Bezymyannyi deposits, except the coarsest-grained and extremely shallow-water sediments, which were not analyzed for their microphytology for obvious reasons. The first six biotas from the middle and lower subformations belong to a single large cycle of shallowing in the history of the basin, when its depth seemingly decreased from 50 to 10 m and less. The variations in biota composition demonstrate two subsequent and distinct trends: initially, a gradual increase of taxonomic diversity and maximum dimensions of microfossils from the first biota to the fourth, and then a decrease of these parameters from the fourth to the sixth biota. For instance, almost all of these biotas contain some, usually very small, amounts of simple sphaeromorphic acritarchs (chiefly *Leiosphaeridea*), colonial coccoidal *Myxococcoides*, and filamentous sheaths of *Eomycetopsis*, and/or *Leiotrichoides*. However, the gigantic acritarchs *Chuaria*, edged *Simia*, and vesicles with internal bodies of *Pterospermopsimorpha* and *Nucellosphaeridium* were found only in the fourth biota, and presumably acantomorphic forms, only in the third and fourth biotas at the respective stratigraphical levels. In a similar manner, the relatively complex pseudofilamentous *Palaeopleurocapsa* and thalluses of *Ostiana* occur only in the fourth biota, which also includes all of the crinkled-costate *Rectia*. As to the overall dimensions of certain morphotypes of the organic-walled microfossils, they significantly increase from the first to the fourth level, and then sharply decrease from the fourth to the sixth level (Fig. 2). In this way, the six biotas of the lower and middle Bezymyannyi subformations clearly show that their most important parameters depend on the sedimentation environments.

The deposits of the upper Bezymyannyi subformation, containing the seventh and the eighth microbiotas, were formed in contrasting environmental settings. The seventh biota originates from deposits corresponding to the maximum shallowing of the late Bezymyannyi basin; while the eighth, from sediments deposited in deeper waters during the subsequent transgression. These two biotas greatly differ in both their taxonomic composition and dimensions of microfossils (Fig. 2), with the eighth biota being very similar in all its parameters to the fourth biota, the richest biota, which formed under comparable conditions. On the contrary, the seventh biota demonstrates a greatly reduced taxonomic composition and contains only small *Leiosphaeridia*, solitary colonial coccoidal *Myxococcoides*, and filamentous sheaths of *Eomycetopsis* and *Leiotrichoides*.

In addition to these taxa, the eighth biota includes the gigantic *Chuaria*, thalluses of *Ostiana*, the crinkled-costate *Rectia*, large trichomes of *Oscillatoriopsis*, and some other forms (Fig. 2).

Thus, all of the Bezymyannyi microbiotas, represented by organic-walled fossils, demonstrate regular paleoenvironmental distribution along the basin profile. The third, fourth, and eighth microbiotas, which are extremely rich, are found confined to deposits of middle depth zones with relatively low hydrodynamic activity. The less diverse first, second, and fifth to seventh microbiotas occur in sediments deposited in deeper (the first and second) or shallower (the fifth through seventh) settings.

Summarizing the data, we may conclude that the most diverse assemblages of the Bezymyannyi organic-walled microfossils, which include morphotypes of complex morphology and maximum dimensions, were restricted in their lateral distribution to the comparatively narrow zone of the basin that corresponded to the proximal part of the outer shelf. The diversity and dimensions of organic-walled microfossils were found to be much lower in the deposits of deeper distal parts of the outer shelf. The same trend is evident when passing to the coastal shallow parts of the Bezymyannyi basin, where the biotas are drastically reduced in their diversity or completely absent.

The microphytological characteristic of the Linok Formation is poorer than that of the Bezymyannyi Formation. Representative microfossils were found only at two levels in its middle dark-colored predominantly shaly subformation, namely within intervals of 15 - 20 and 30 - 40 m from the base, in one section near the mouth of Gremyachii Creek, a tributary of the Nizhnyaya Tunguska River. Other levels of this and other sections of the subformation contain only solitary small acritarchs and poorly preserved fragments of filamentous forms. Both relatively rich Linok biotas are confined to intervals of the subformation that were abundant in shaly siltstone interbeds.

The relatively diverse Linok microbiotas are undoubtedly peculiar in comparison to those from the Bezymyannyi Formation. The Linok organic-walled microbiota is dominated by two morphological groups, namely acritarchs and intricate forms (58 - 60% and 27 - 29% of all encountered forms, respectively). Acritarchs are represented here by relatively simple vesicles of *Leiosphaeridia*, *Chuaria*, and by solitary large *Nucellosphaeridium* forms. Their diameters vary from 30 to 500 μm, sometimes up to 550 - 570 μm with two dimension peaks at 120 - 150 and 350 - 400 μm. The examined sample does not contain coccoidal forms, except for several groups of simple coccoidal cells. Filamentous microfossils are represented by rare fragments of *Oscillatoriopsis* trichomes (6 - 10 μm wide) and equally rare *Leiotrichoides* sheaths (8 - 10 and 20 - 30 μm wide). There are more numerous peculiar sheaths of *Siphonophycus* with transverse structural elements (20 - 22 and 40 μm wide), and tubular forms similar to *Rectia*, but with a less developed sculpture. The Linok microfossils of complicated form are represented by very few specimens of rounded envelopes with long processes. Inside the latter, parallel to their long axis, there is a well expressed dark granulated zone (the envelope diameter is 30 - 50 μm; the processes are 18 - 20 or 40 μm wide, and up to 100 μm long; the dark zone is 14 - 16 μm wide). These fossils are similar to the germinating structures of some ancient fungi: *Germinosphaera*, *Caudosphaera* (*Mikrofossilii* ..., 1989) or *Vendomyces* (Burzin, 1992). Another significant element of the Linok complicated microfossils is represented by dark ribbonlike thalluses with occasional signs of branching. The thalluses are 20 - 25 and 35 - 40 μm wide. The most specific among them are the large (up to 80 - 90 μm wide) segmented trichomelike forms, with cylindrical or rounded segments ("cells"). The biological affinity of such microfossils is unclear. The difference between the two Linok organic-walled microbiotas is the nearly complete absence of complicated forms and the impoverished set of acritarchs in the second assemblage (their maximum size is 350 μm).

The Sukhaya Tunguska Formation, consisting of purely carbonate deposits, is free of organic-walled microfossils.[1] However, its chert nodules displayed numerous well preserved silicified microfossils (Mendelson and Schopf, 1982; Golovenok and Belova, 1993; Sergeev, personal communication), which constitute a rather representative assemblage. In addition to the usual Precambrian coccoidal and filamentous *Gloeodiniopsis*, *Eosynechococcus*, *Eoaphanocapsa*, *Myxococcoides*, *Eomycetopsis*, *Siphonophycus*, *Oscillatoriopsis*, and some other taxa, there are specific *Eogloeocapsa* colonies yielding packets of cells, *Coniunctiophycus* colonies of complex structure, multiserial *Eomicrocoleus* and stalklike cyanobacteria *Polybessurus*. Occurrences of presumably acantomorphic acritarchs *Eomycrhystridium* (?) and simple *Leiosphaeridia* should be especially noted.

Among the above forms, those widespread throughout the entire section are only the coccoidal *Eogloeocapsa* and filamentous *Eomycetopsis* and *Siphonophycus*. The other taxa are confined to two different stratigraphical levels. One of them is in the middle part of the formation (near the top of its lower subformation, according to the authors' scheme, or at the base of its upper subformation, according to V.K. Golovenok). The second level is located near the top of the upper subformation. The lower level yields more diverse microbiota. In addition to the above coccoidal and filamentous taxa, it contains acritarchs *Eomicrhystridium* (?) and *Leiosphaeridia*. This lower assemblage of the Sukhaya Tunguska microbiota is found in deposits that accumulated under shallow and middle depths within the zone of storm and less frequent fair-weather wave activity. The less representative upper microbiota was found in much shallower

[1] The references on rare sphaeromorphids from that formation (Timofeev *et al.*, 1976) require critical revision.

deposits (including those with traces of subaerial exposition). Thus, the mineralized Sukhaya Tunguska microfossils demonstrate that their taxonomic diversity depends on environmental settings, as in the case of the Bezymyannyi organic-walled fossils. Both the Sukhaya Tunguska and the Bezymyannyi microbiotas proved to be more diverse and rich in specific morphotypes, when their sedimentation settings changed from the shallowest to relatively deeper zones.

The general sequence of the organic-walled and silicified microfossils from the Bezymyannyi, Linok, and Sukhaya Tunguska formations did not exhibit any variation trend in their taxonomic composition, which could be attributed to the general evolution of the assemblages and their members. At the same time, the observed taxonomic variations can be directly correlated with the environmental peculiarities of the enclosing deposits.

In general, the deposits containing the above microfossils demonstrate a wide range of basinal environments: from the deep distal parts of the shelf to the near-shore zones of its interior. Within this environmental range, the Bezymyannyi and Linok microbiotas form four groupings. Each of them has its own specific taxonomic composition, definite dimensional parameters of transitional morphotypes present in two or several groupings, and confinement to certain bathymetric zones of the basin, and hence to a particular range of hydrodynamic conditions. We can conventionally name these groupings as: shallow, transitional, "optimal" (medium-deep), and deep-water. These groupings are represented by organic-walled and silicified forms that originated respectively from basins of siliciclastic and carbonate sedimentation, and essentially differ from one another in both their taxonomic composition and the ecological confinement of all their leading components. Among the microfossils considered above, the shallow grouping embraces the second, sixth, and seventh biotas from the Bezymyannyi Formation. The transitional grouping comprises the third and fifth Bezymyannyi biotas, while the "optimal" grouping consists of the fourth and eighth Bezymyannyi biotas. The deep-water grouping is composed of the first Bezymyannyi and the ninth to tenth Linok biotas. Now we shall summarize the most characteristic taxonomic features of the groupings.

The shallow grouping of organic-walled microfossils was formed within the limits of the Bezymyannyi inner shelf. It has three specific features: (1) it has a very low diversity; (2) it contains only such transitional Proterozoic taxa as *Leiosphaeridia*, *Myxococcoides*, *Eomycetopsis*, and *Leiotrichoides*; and (3) all morphotypes are represented only by the smallest forms (Fig. 3). The combination of acritarchs, dominated by strongly corroded and mechanically disturbed incomplete vesicles, with very small fragments of coccoidal colonies and sheath fragments suggests that the assemblage is prevailed by allochthonous elements of the Precambrian biota.

The transitional grouping of organic-walled microfossils confined to the relatively deeper parts of the inner shelf seems to interconnect the poorest shallow-water biota with the richest and representative "optimal" one. The biotas of the transitional grouping are essentially enriched in acritarchs (almost completely represented by *Leiosphaeridia*) and several varieties of colonial coccoidal forms (chiefly *Myxococcoides*) with obviously enlarged dimensions.

The "optimal" grouping was discovered in the deposits of the proximal gently sloping part of Bezymyannyi outer shelf. The relevant microbiotas differ from all other samples in their high taxonomic diversity, in the presence of very large specimens of all morphological varieties (Figs. 2 and the table), and in the appearance of some indicative taxa typical only of this grouping. In our materials, the latter are represented by acritarchs *Chuaria* and *Simia*, and also by presumably acantomorphic forms associated with elongated thalluses of coccoidal *Ostiana*, filamentous smooth-walled *Asperatofilum* and *Taenitrichoides*, crinkled-costate *Rectia*, multiserial *Polytrichoides*, and peculiar *Oscillatoriopsis* trichomas. Occurrence of the two latter taxa only in the eighth biota is the sole difference between the two biotas assigned to this grouping.

The deep-water grouping of organic-walled microfossils occurs in deposits of distal parts of the outer shelf. This grouping is marked by the occurrence of medium-sized *Leiosphaeridia* and *Chuaria* along with dark, structureless thalluses, including those with longitudinal sculpture, and some other forms probably formed from fungi (vesicles with processes of different length). There are also sheath fragments of *Rectia*, particularly widespread in the medium-deep zone and seemingly representing allochthonous elements in the deep-water settings.

Thus, the data on the Bezymyannyi and Linok formations demonstrate that in transition from the peritidal siliciclastic zones of the basin towards deeper settings, the microbiotas rapidly and sharply change their taxonomic composition, overall dimensions, and ecological character. The poorest and most inexpressive allochthonous grouping, enclosed in the shallow-water siliciclastic deposits, gives way, passing through the relatively narrow transitional zone, to the taxonomically richest and most diverse, chiefly benthic, "optimal" grouping of organic-walled microfossils characteristic of the medium-depth environments. Further, in the offshore parts of the basin, the taxonomic diversity of the biotas decreases, and some highly specific benthic forms appear there.

Following from the above, an important feature of four Turukhansk microfossil groupings is a particular proportion of planktonic and benthic forms in them (when characterizing assemblages of these forms, one should distinguish the quantitative and taxonomic aspects). Despite all the objective difficulties in discriminating the plankton and benthic assemblages of organic-walled microfossils, we agree with some scientists (e.g., Vidal and Knoll, 1983; Burzin, 1993; Vidal

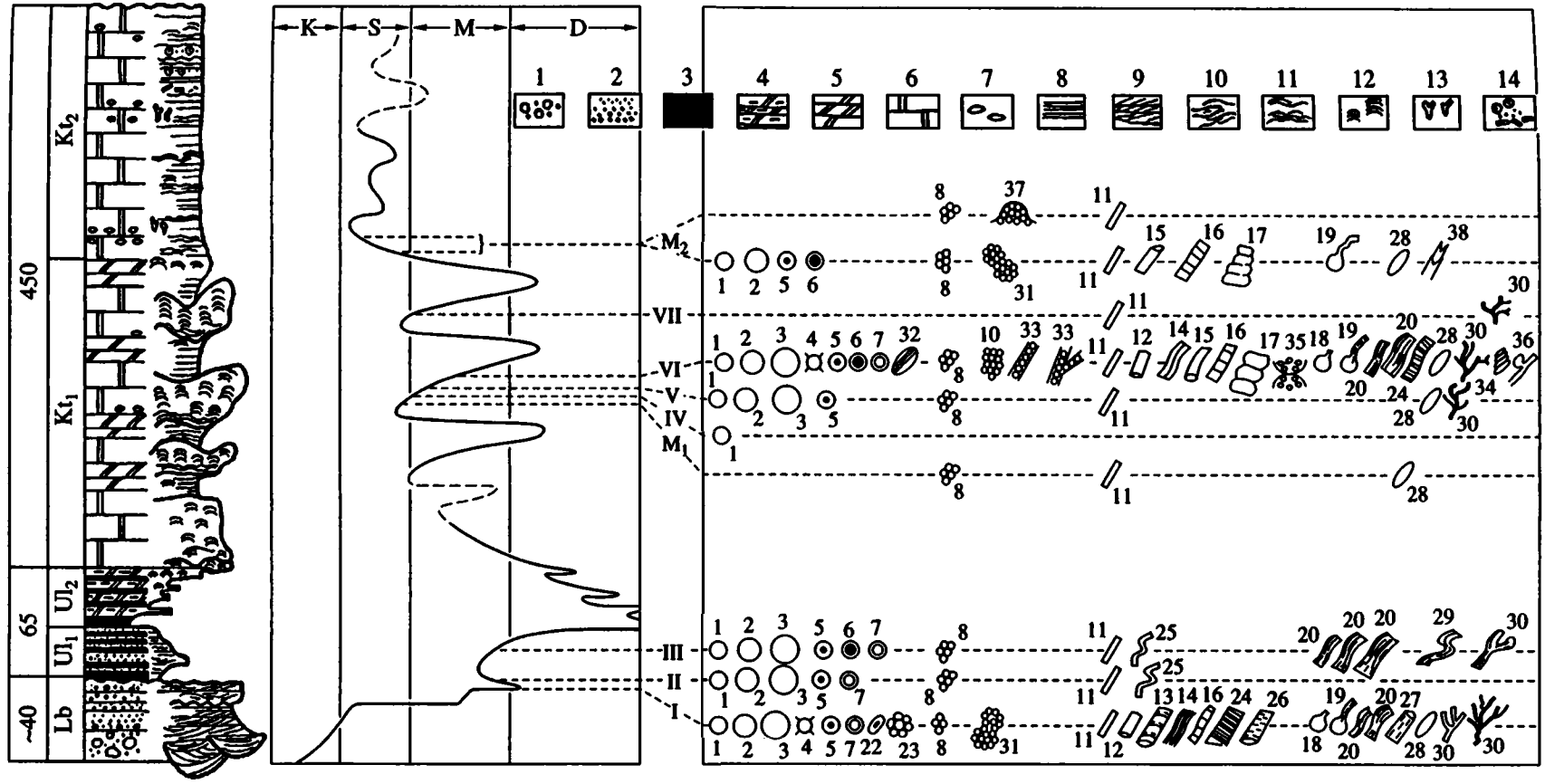

Fig. 3. Distribution of microfossils in the Riphean deposits of the Anabar massif and the environmental variations during the sedimentation of the formations: Labaztakh (Lb), Ust'-Il'ya (Ul), and Kotuikan (Kt).

K – terrestrial environments. Marine environments: S – shallow-water; M – medium-deep; D – deep-water. Composition of deposits: 1 – gravelstones and sandstones; 2 – coarse- to fine-grained sandstones; 3 – shales; 4 – dolomitic marl; 5 – clayey dolomites; 6 – dolomites; 7 – chert nodules; 8 – thin horizontal lamination; 9 – cross bedding; 10 – cross-wavy bedding; 11 – stratiform and columnar-stratiform stromatolites; 12 – columnar stromatolites; 13 – microstromatolites (of the *Asperia* type); 14 – carbonate clastic rocks (doloarenites, flat-pebbled conglomerates, and oolites). Morphotypes of organic-walled and silicified microfossils: 1 – *Leiosphaeridia* (the diameter up to 100 μm), 2 – *Leiosphaeridia* and *Chuaria* (the diameter up to 500 μm), 3 – *Chuaria* (the diameter 700 - 1000 μm and more), 4 – presumably acantomorphic acritarchs, 5 – *Nucellosphaeridium*, 6 – *Pterospermopsimorpha*, 7 – *Simia*, 8 – *Myxococcoides*, and other relatively simple coccoidal forms, 9 – *Pleurocapsa*-like aggregates of coccoidal cells, 10 – *Ostiana*, 11 – *Eomycetopsis* and/or *Leiotrichoides*, 12 – *Asperatofilum* and/or *Taenitrichoides*, 13 – *Rectia*, 14 – *Polytrichoides*, 15 – *Siphonophycus*, 16 – *Oscillatoriopsis* (*Cephalophytarion*, *Palaeolyngbya*) and other relatively simple trichomes, 17 – large trichomelike forms, 18 – reproductive structures of algae and/or of fungi, 19 – fungilike (hyphomicobal) fossils, 20 – nonbranching ribbonlike thalluses, 21 – *Polybessurus*, 22 – *Konderia*, 23 – *Satka*, 24 – *Botuoba*-like trichomes, 25 – *Heliconema*, 26 – *Plicatidium*, 27 – *Eomicrocoleus*, 28 – *Eosynechococcus*, 29 – spirally curved thalluses, 30 – branching thalluses (including those close to *Ulophyton*, *Archaeoclada*), 31 – *Coniunctiophycus* and/or *Eomicrocystis*, 32 – *Cucumiforma*, 33 – chainlike aggregates of coccoidal cells inside cylindrical sheaths (including those with branching structure), 34 – *Elatera*, 35 – *Eosolena*, 36 – *Vanavarataenia*, 37 – *Eoentophysalis*, 38 – *Circumvaginalis*. I - VII – occurrence levels of organic-walled microfossils; M_1, M_2 – occurrence levels of mineralized microfossils.

and Nystuen, 1990) that it is still possible, at least for some indicative morphotypes (acritarchs, mats of coccoidal and filamentous forms, metaphytic thalluses, and others). Identification of this kind is based on actualism principles, common sense, and experience in ecological interpretation of silicified forms.

From this position, one can say that the above groupings of organic-walled microfossils sharply differ in their ecological composition. Judging by the composition of their characteristic morphotypes, namely the acritarchs, rare colonial coccoidal and filamentous forms, the shallow-water and transitional groupings were completely or almost completely dominated by allochthonous (planktonic?) elements. The other two groupings, the "optimal" and deep-water, had a more complicated ecological structure. All members of the "optimal" group, except acritarchs and some fragmented coccoidal and filamentous forms, belonged to the benthic assemblage. Taxa of the latter were more abundant in quantity (over 70% of all genera and 60% of all species), comprising 50% of all organic-walled microfossils found there. As for the deep-water grouping, its most abundant planktonic acritarchs are associated with more diversified taxa of benthic filamentous and other microfossils of complicated structure.

The shallow-water grouping of silicified microfossils, as was mentioned before, is represented by two Sukhaya Tunguska biotas. They were formed in peritidal settings of carbonate platform. Both biotas are dominated mainly by allochthonous microfossils, while distinct benthic assemblages of coccoidal and/or filamentous forms are almost completely absent (this supports the conclusion that subaquatic settings were characteristic of the Sukhaya Tunguska time). Still, the above biotas contain some taxa of definite ecological settings, first of all, the benthic stalklike cyanobacteria and planktonic acritarchs.

The Anabar Massif

The sequence of the Labaztakh, Ust'-Il'ya, and Kotuikan deposits corresponds to a large transgressive cycle, which can be compared in structure with that of the Bezymyannyi, Linok, and Sukhaya Tunguska deposits (Fig. 1). The upper part of the Labaztakh Formation (40 m thick) starts the above transgressive cycle, and consists of two members. The lower member (10 m thick) is composed by variously sorted, inequigranular oligomictic sandstones with large cross-bedding and rare gravel accumulations. The rocks of this member show systems of broad (up to 1 m in width) troughs filled with coarse-grained sandy material. The lowest parts of these troughs contain accumulations of gravel and well-rounded cobbles. Structurally, these rocks correspond to the alluvium of large plain rivers (Zlobin and Golovanov, 1970).

The upper member of the Labaztakh Formation (30 m) consists chiefly of fine- to medium-grained sandstones and rare gravelstones. The latter rapidly decrease in abundance upward in the section, giving way to occasional members of sandstones with regular planar cross-bedding. The amplitude of the latter is considerably lower than that of similar rocks from the lower member. These depositional features show that the upper member was formed on an alluvial coastal-marine plain.

The succession of the upper member is terminated by the thick-bedded member of medium- to coarse-grained sandstones (3 m thick) containing glauconite. These sandstones show tabular to wavy and hummocky cross-bedding. The essential oscillation amplitudes (up to 20 cm) in the cross-bedded units testify to the intensive wave activity within the sedimentation setting. This fact, as well as the presence of glauconite in the sandstones, suggests that depositional settings were open-marine in character.

The overlying Ust'-Il'ya Formation (60 - 70 m thick) is separated by a sharp boundary from the Labaztakh cross-bedded sandstones. In terms of its composition, the Ust'-Il'ya Formation takes an intermediate position between the siliciclastic Mukun and essentially carbonate Billyakh groups of the Anabar Riphean sequence (Fig. 1). This formation is subdivided into two subformations. The lower one (30 - 35 m thick) consists of frequently alternating dark-grey shales, siltstones, and fine-grained sandstones of lighter color. The shales and siltstones are of even and fine horizontal lamination. The continuous sandstone beds of 1 - 5 cm to 2 m thick are horizontally bedded (with occasional graded structure), which, most likely, can be related to deposition of sandy material from suspension flows. The upper parts of such beds reveal internal local scours and various patterns of small-scale ripple-lamination resulting from the activity of unidirectional bottom currents. Such depositional features are most characteristic of the distal tempestites in remote parts of the shelf (Aigner, 1985). The thickness of the sandy layers gradually decreases upwards, probably indicating that the sedimentation zone was shifted to a more distal and presumably deeper part of the shelf. On the whole, this subformation can be regarded as a unit formed below the storm wave base.

The upper Ust'-Il'ya subformation (30 m thick) is composed of dark-grey and dark-green shales; in its upper part, there are also clayey dolomites and marls increasing in abundance upward in the section. At the base of this subformation, there is a 3 m-thick member of characteristic dark-green horizontally bedded shales, which are almost completely devoid of any siltstone admixtures. In the upper part of this subformation, the amount of the carbonate material gradually increases, and some structures testifying to the strengthening hydrodynamic activity also appear. Among dolomites occurring here, one can see the increasing role of stromatolite varieties with columnar (*Kussiella kussiensis* Kryl.) and stratiform (*Stratifera*) buildups (Komar, 1966). The latter are confined to more clayey rocks, while the columnar morphs build small bioherms and biostromes closely associated with detrital varieties of carbonate sediments. In the uppermost part of the Ust'-Il'ya Forma-

Dimensions of different morphotypes of organic-walled microfossils from Riphean deposits of the Turukhansk uplift (the Bezymyannyi and Linok formations) and from the Anabar massif (the Ust' Il'ya and Kotuikan formations)

Formation	Subformation	Occurrence level of organic-walled microfossils	Morphotypes of microfossils						
			acritarchs	colonial coccoidal forms	filamentous forms: sheaths	filamentous forms: trichomes	filamentous forms: trichomelike bodies	thalluses: nonbranching	thalluses: branching
THE TURUKHANSK UPLIFT									
Linok	Upper								
	Middle	X	30 - 350 (120 - 150)	8 - 12	8 - 40	6 - 10		20 - 35	
		IX	30 - 570 (120 - 150, 350 - 400)	8 - 12	8 - 30	6 - 10	80 - 90	20 - 40	
Bezymyannyi	Upper	VIII	10 - 750 (150 - 500)	16 - 24	4 - 70 (4 - 30)	10 - 48			
		VII	20 - 50	10 - 12	2 - 12				
	Middle	VI	10 - 250 (30 - 50)	10 - 25	6 - 12				
		V	20 - 500 (100 - 150)	8 - 30	6 - 12				
		IV	10 - 1000 (250 - 500)	8 - 50 (10 - 30)	4 - 70 (4 - 20)				
		III	10 - 550 (250 - 450)	12 - 30	4 - 12				
		II	10 - 95 (10 - 40)	18 - 20	4 - 8				
	Lower	I	8 - 350	10 - 12	4 - 12			64 - 90	
THE ANABAR MASSIF									
Kotuikan	Lower	VII			2 - 20			10 - 20	10 - 20
		VI	10 - 1400 (150, 300, 500 - 700)	10 - 40 (20 - 25, 30 - 40)	2 - 100 (2 - 4, 8 - 14, 20 - 40)	8 - 30	120 - 260 (120 - 160)	16 - 160 (20 - 30, 40 - 60, 100 - 140)	10 - 40
		V	18 - 640 (100 - 300)	16 - 18	4 - 10			10 - 40 (10 - 20, 30 - 40)	10 - 20
		IV	40 - 50						
Ust'-Il'ya	Lower	III	30 - 1500 (70 - 500)	16 - 18	2 - 14			10 - 160 (10 - 12, 16 - 18)	10 - 20
		II	10 - 1500 (150 - 300)	16 - 18	2 - 14			20 - 90 (80)	
		I	12 - 750 (150 - 300)	14 - 20	2 - 140 (8 - 14, 60 - 70)	4 - 30		10 - 60 (10 - 12, 30, 50)	40 - 100

Note: All dimensions are given in microns. The parentheses indicate the dimensions of prevailing forms. For colonial coccoidal forms the diameter of individual cells is indicated; for thalluses and filamentous forms (sheaths and trichomes), their width.

tion, the amount and thickness of the shale interbeds drastically decreases; their continuous layers disappear to fill only grooves between bioherms.

The described upper Ust'-Il'ya sequence corresponds to the stage when the basin became shallow. On the whole, the entire Ust'-Il'ya Formation corresponds to two related and symmetric stages of deepening and shallowing (Fig. 3). The basin reached its maximum depths during the deposition time of the green shale member at the base of the upper subformation.

The Kotuikan Formation, which is about 500 m thick, conformably rests on the Ust'-Il'ya deposits, and is disconformably overlain by the Yusmastakh Formation (Fig. 1). It is divided into two subformations (Fig. 3), the lower of which consists of two members. The lower member (40 m thick) consists of light-grey stromatolitic dolomites with their clayey varieties and subordinate shales in the upper part. The stromatolites form persistent biostromes built by columnar branching, nonbranching, and rare nodular and stratiform morphs (Komar, 1966; Golovanov, 1970; Krylov, 1975).

The upper member (about 250 m thick) is an intricate biohermal complex (Serebryakov, 1975) dominated by large stromatolitic bioherms up to 20 - 30 m thick and extending over several hundred meters. The bioherms are confined to separate intervals of the member section, being separated by interbiohermal clay-carbonate deposits. Together with the latter, they are overlain by persistent members of horizontally bedded clayey dolomites. The bioherms are underlain by light-grey dolomites of varying thickness with indistinct stratiform stromatolites, tepee structures, and concretions of early diagenetic cherts.

Within the bioherms of vertical change in the stromatolite morphology, there is a general trend from stratiform rarer columnar-layered and nodular stromatolites at the base of bioherms to columnar branching *Kussiella* in their middle, thickest part, and then again to stratiform morphs near the top. Sometimes, columnar nonbranching stromatolites with conical laminae, described as *Conophyton cylindricus* Masl. (Golovanov, 1970), appear between horizons of *Kussiella* and *Stratifera*. The intercolumnar spaces of *Kussiella* usually display alternations of clastic fine-grained carbonates and flakestones persistent along the strike for at least several meters. The flakestones disappear between the columns of conophytes. The stromatolitic dolomites within the bioherms compose persistent strata 10 to 40 cm in thick, separated by erosional surfaces corresponding to pauses in the growth of a bioherm dome.

The interbiohermal clay-carbonate facies differ from one level to another. At the levels synchronous to the lower parts of the bioherms, they demonstrate certain crack structures on the bedding surfaces, probably the result of the subaerial desiccation of the sediment. At the levels corresponding to the middle and upper horizons of the bioherms, the interbiohermal space is filled with coarse-detrital carbonate rocks, and above them, with fine-grained dolomitic sandstones and siltstones having cross-wavy and occasionally hummocky cross-bedding. Upsection, the role of clastic carbonates gradually decreases, while fine-grained and clayey carbonates become more abundant. Still higher, there are members of dark clayey dolomites mentioned above. They reveal a fine horizontal bedding, formed by interrupted films of organic matter. Judging by the facies relation, the relief in bioherms, life-time varied from several to 10 - 15 meters.

The upper subformation of the Kotuikan Formation (about 200 m thick) is a sequence of light-grey dolomites, at some levels containing early diagenetic chert nodules rich in silicified microfossils (Golovenok and Belova, 1981, 1984; Yakshin, 1991; Sergeev, 1993). The composition of the subformation is dominated by stromatolitic dolomites, massive micritic dolomites with horizontal lamination, and clastic dolomites. In the upper part of the subformation, there are subordinate interbeds of fine- to medium-grained dolomitic sandstones. The stromatolites are mainly stratiform varieties with a very low relief. There are also microstromatolites (*Microstylus*, *Asperia*), though they occur much more seldom.

Now we shall consider the environmental settings of the Kotuikan deposits. The lower member of the Kotuikan Formation and the upper Ust'-Il'ya Subformation show a common succession of changes in the paleoenvironments and type of sedimentation, from clayey-siliciclastic deep-water to the shallow-water carbonate stromatolitic one. This part of the Anabar Riphean sequence can be considered to reflect a large cycle of the basin shallowing. The upper member of the lower Kotuikan Subformation reveals at least two relatively large transgressive-regressive cycles. The transgressive part of the cycles is sufficiently well represented in the section, the regressive part is strongly reduced.

The cycle begins with the growth of the biohermal domes of stratiform stromatolites. During this time, the surface of the growing domes was influenced by frequent sea level oscillations, including moments of their emergence marked by the dissecation cracks in the lower horizons of the interbiohermal sediments. When flooded, the bioherms resumed their growth and the interbiohermal depressions accumulated carbonate-clay sediments.

As the transgression proceeded and the subaquatic conditions became permanently dominant, the interruptions in the growth of the biohermal domes were controlled by changing hydrodynamic conditions that, in their turn, were influenced by short-term sea level oscillations. The high-energy environments prograded into interbiohermal depressions, where detrital trains and wavy oscillation structures, including hummocky lamination in fine-grained sediments, were formed.

With the further advance of the transgression, the depth growth, and decline of the general agitation of the bottom water, carbonate-clay sediments continued to

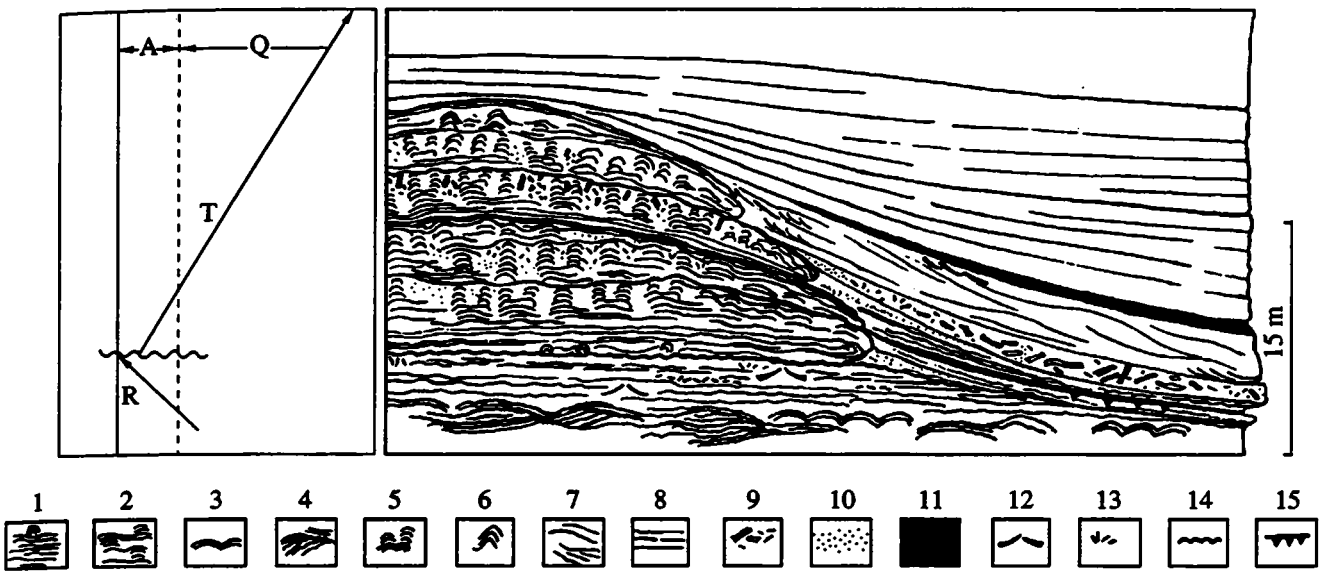

Fig. 4. Structure and sequence of biohermal and interbiohermal facies in the transgressive-regressive cycle from the lower Kotuikan Subformation (also see Fig. 3).

Lithological characteristics of deposits: 1 - 6 – stromatolites: 1 – stratiform and nodular; 2 – columnar-stratiform (of the *Omachtenia* type); 3 – linked stromatolitic domes, frequently elongated in one direction; 4 – large low-relief domes with scour overgrowth structures; 5 – columnar branching buildups (*Kussiella*); 6 – columnar nonbranching buildups (*Conophyton*); 7 - 13 – mechanogenic deposits: 7 – dolomites with hummocky and swally cross stratification; 8 – clayey dolomites with horizontal lamination; 9 – chaotically oriented flat-pebble conglomerates; 10 – fine-grained dolomitic sandstones; 11 – interbeds of shales; 12 – tepee structures; 13 – edgewise flat-pebble conglomerates; 14 – erosional surfaces; 15 – desiccation cracks. T – transgressive part of the cycle; R – regressive part of the cycle; A – settings with periodical subaerial exposures; Q – subaqueous settings.

level the biohermal relief. Under such conditions, the clay material was strongly mixed with carbonate silt. The distribution of sediments at that time was mainly dependent on echos from high-energy events. The moments of highest sea-level position are recorded by the deposition of clay-carbonate shales overlapping the bioherms (Fig. 4). In the middle phase of the transgression, the depths of the early Kotuikan basin were at least 10 to 15 m, as testified by the relief bioherms' life-time. At the time of high sea level, the depths probably reached several tens of meters. As for the highest horizons of the clay-carbonate members overlying the bioherms-bearing unit, and dolomite members underlying them, their position in the sequence and characteristic structures indicate the regressive phase of the cycle under discussion. Among the lower Kotuikan deposits, these sediments were formed under the shallowest conditions.

The upper Kotuikan deposits were accumulated even in the shallower environments, as compared with the lower Kotuikan time. The upper subformation is dominated by tidal-flat deposits. Among them, the most numerous are sediments of extremely shallow settings and ephemeral basins. Against this background, the uppermost part of the subformation includes a relatively greater amount of deep-water deposits, namely coarse varieties of clastic dolomites and dolomitic sandstones (Fig. 3).

Considering the taxonomic composition and paleoenvironmental confinement of the Anabar microfossils, we have to note that, in general, their spatial distribution repeats that inferred for the Turukhansk uplift. Rich microbiotas here, like in sections there, were found only at some discrete levels, bearing about 90% of all discovered forms. Beyond these levels, all studied samples contain either drastically impoverished biotas, devoid of any characteristic morphotypes, or many films of decayed organic substance, absolutely lacking identifiable microfossils.

The general characteristic of the Kotuikan and Ust'-Il'ya organic-walled microfossils, described by Veis and Vorob'ieva (1992), is now widened, because several new taxa were discovered in former and also new, recently found occurrences about 130 m above the Kotuikan Formation base, in the locality on the left bank of Kotuikan River, near the mouth of the Ulakhan-Tomulakh Creek (Plates II and III).

In the Ust'-Il'ya Formation, all of the organic-walled microfossils found are confined to the lower subformation. They originate from several localities situated on the Kotuikan River, 2.5 km above the mouth of the Il'ya River and in its circum-mouth area. The bulk of the microfossils were discovered here at three stratigraphical levels: 2 - 4.5, 7 - 8, and 21 - 23 m above the formation base. The assemblage of these microfossils embraces all of the main Precambrian morphotypes, representing over 20 genera and 30 species described in the literature. In addition, it includes a number of microfossils that cannot be attributed to

any of the presently accepted taxa; their place in the system of organic-walled microfossils requires particular attention. On the whole, this association is dominated by acritarchs (about 50% of all forms studied) and by complex forms (30%), including several kinds of thalluses, some of them branching. Filamentous and colonial coccoidal forms occur also in subordinate amounts (not more than 15% and 5%, respectively). The fossils are highly concentrated in samples being well preserved and have a generally complex structure and large dimensions common to most morphotypes. The acritarch group is dominated by the *Leiosphaeridia* and *Chuaria*. Over half of these forms have dimensions exceeding 200 and 500 μm respectively, and some individual chuariamorphs reach 1500 μm. In addition, the Ust'-Il'ya acritarchs display vesicles with internal bodies, i.e., the *Konderia*, *Nucellosphaeridium*, *Pterospermopsimorpha*, and some extremely large forms (up to 460 - 480 μm in diameter). Together with them, there are *Simia* (up to 260 - 280 μm), smaller tablelike *Satka* (50 μm), and presumably acantomorphic forms, including membrane-rimmed representatives (from 300 to 320 μm).

The thalluses from the Ust'-Il'ya Formation display a wide dimensional range from 10 to 160 μm. There are four main morphological varieties here: (1) the simplest ribbonlike forms of a wide dimensional range (the predominant thalluses are 10 - 12, 16 - 18, 30, 50 and 80 μm in width); (2) branching thalluses, among which the most distinguishable are bushlike forms similar to *Ulophyton* (the bushes reach 1500 μm in height, their branches at the base are up to 100 μm wide), and more complex bodies swollen (or not swollen) at the ends of the branches (the dimensional peaks are 10 - 12, 16 - 18, 40 and 50 μm; the largest forms were previously assigned to *Archaeoclada*); (3) planospiral forms with a dark longitudinal zone of 40 μm wide; and (4) thin filmy forms with coarse transverse ribs, producing a characteristic rhombic pattern when two or several forms are superimposed (the fragments are up to 120 - 140 μm wide). The group of complex organic-walled microfossils may possibly comprise vesicles with processes, which are similar to *Germinosphaera*, *Caudosphaera*, and/or *Vendomyces* forms (the diameter of the spheral body is 30 - 110 μm), and also large tabular segmented *Eosolena* specimens.

The filamentous organic-walled microfossils from the Ust'-Il'ya Formation are represented by sheaths and trichomes. The former can be subdivided into 5 types: (1) smooth-walled forms 2 - 4 to 140 μm wide, which can be assigned to *Eomycetopsis*, *Leiotrichoides*, *Asperatofilum*, and *Taenitrichoides*; (2) planospiral *Heliconema* (10 - 12 and 18 - 20 μm wide); 3) tubular forms with different transverse sculpture and dimensions: *Rectia* with crinkled-costate sculpture and 45 - 50 μm wide; *Plicatidium* with fine, nonpersistent sculpture and 76 - 140 μm wide; with frequent transverse striation and 12 - 14 to 34 - 36 μm wide; (4) multifilamentous sheaths of *Eomicrocoleus* and *Polytrichoides*; and (5) large ellipsoidal *Eosynechococcus* vesicles (up to 100 μm wide). The trichomes and trichomelike microfossils are represented by two varieties: *Oscillatoriopsis* forms with short cylindrical cells (up to 20 - 30 μm in diameter) and *Palaeolyngbya* trichomes of similar structure, enclosed in a well developed smooth-walled sheath (the trichome is 10 and the sheath is 20 μm wide).

The colonial coccoidal forms in this association are represented by: *Myxococcoides* (the cell diameter is 14 and 20 μm), sometimes with very distinct and dense cellular walls; aggregates of subspheral *Coniunctiophycus* colonies (the cell diameter in the colonies is 18 - 20 μm); and coenobial accumulations of acritarch-

Plate II. Microfossils from the Ust'-Il'ya and Kotuikan Upper Precambrian formations on the western slope of the Anabar massif.

All of the described specimens are stored at the Geological Institute of the Russian Academy of Sciences, Moscow.

1- 3, 5 - 7, 13, 15, 17, 18, 20 - 23, 26 – the Ust'-Il'ya Formation, the lower subformation; 3, 5 - 7, 13, 15, 17, 18, 20, 26 – the first occurrence level of organic-walled microfossils (the location on the right bank of the Kotuikan River, 2.5 km upstream the mouth of the Il'ya River); 1, 2, 23 – the second occurrence level of silicified microfossils, the same location; 21, 22 – the third occurrence level of silicified microfossils (the left bank of the Il'ya River, near its mouth); 4, 8 - 12, 14, 16, 19, 24, 25, 27 – the Kotuikan Formation, the lower subformation, a new location; the left bank of the Kotuikan River, opposite the mouth of the Ulakhan-Tongulakh Creek.

All of the described specimens belong to the optimal grouping of Anabar organic-walled microfossils.

Bar for 1, 2, 14, 21 - 23 = 350 μm; for 3, 10 = 800 μm; for 4, 8, 9, 11, 12, 26 = 400 μm; for 5 = 500 μm; for 6, 13 = 220 μm; for 7 = 80 μm; for 15 = 530 μm; for 16 - 19, 25, 27 = 120 μm; for 20 = 1000 μm; for 24 = 280 μm.

1, 2, 5 – *Nucellosphaeridium nordium* (Tim.): 1 – sp. no. 191/1550, 2 – sp. no. 191/1551, and 5 – sp. no. 191/1510; 3, 4 – *Chuaria circularis* Walcott emend. Vidal et Ford: 3 – sp. no. 191/62 and 4 – sp. no. 591/ 952; 6, 9 – *Myxococcoides grandis* Horod. et Donald.: 6 – sp. no. 191/653 and 9 – sp. no. 591/954; 7 – *Myxococcoides reticulata* Schopf; sp. no. 191/612; 8 – *Myxococcoides inornata* Schopf; sp. no. 591/914; 10 – fragment of a broad thallus with transverse structure; sp. no. 591/955; 11 – *Ostiana microcystis* Herm.; sp. no. 591/956; 12 – fragment of mat composed by sheaths *Leiotrichoides typicus* Herm.; sp. no. 591/957; 13 – coenobial aggregate of small vesicles *Leiosphaeridia* sp. no. 191/618; 14, 20 – fragments of intricately branching thalluses *Ulophyton ripheicum* Tim. et Herm.: 14 – sp. no. 591/958 and 20 – sp. no. 191/362; 15 – ring-shaped fossil with a process; sp. no. 191/359; 16 – chainlike aggregate of subspherical cells in a cylindrical sheath; sp. no. 591/960; 17 – *Oscillatoriopsis media* Mendelson et Schopf; sp. no. 191/610; 18 – a large thin-walled tubular form with distinct transverse striation yielding a typical rhombic pattern when specimens are overlapped; sp. no. 191/690; 19 – fragment of a filamentous sheath with distinct transverse structure reflecting the cellular composition of trichome (the latter is absent); sp. no. 590/861; 21 – fragment of a broad thallus, bifurcating and thinning towards its end; sp. no. 491/462; 22, 23 – subspiral filamentous forms: 22 – the fossil comprises many interwoven filaments; sp. no. 491/464 and 23 – the fossil comprises two filaments, sp. no. 191/1565; 24 – a broad sheath with a transverse structure and remnants of collapsed trichome (? *Palaeolyngbya sphaerocephala* Herm. et Pylina); sp. no. 591/966; 25, 27 – *Polytrichoides lineatus* Herm.: 25 – sp. no. 591/667 and 27 – sp. no. 591/368; 26 – *Asperatofilum experatus* (Herm.); sp. no. 191/668.

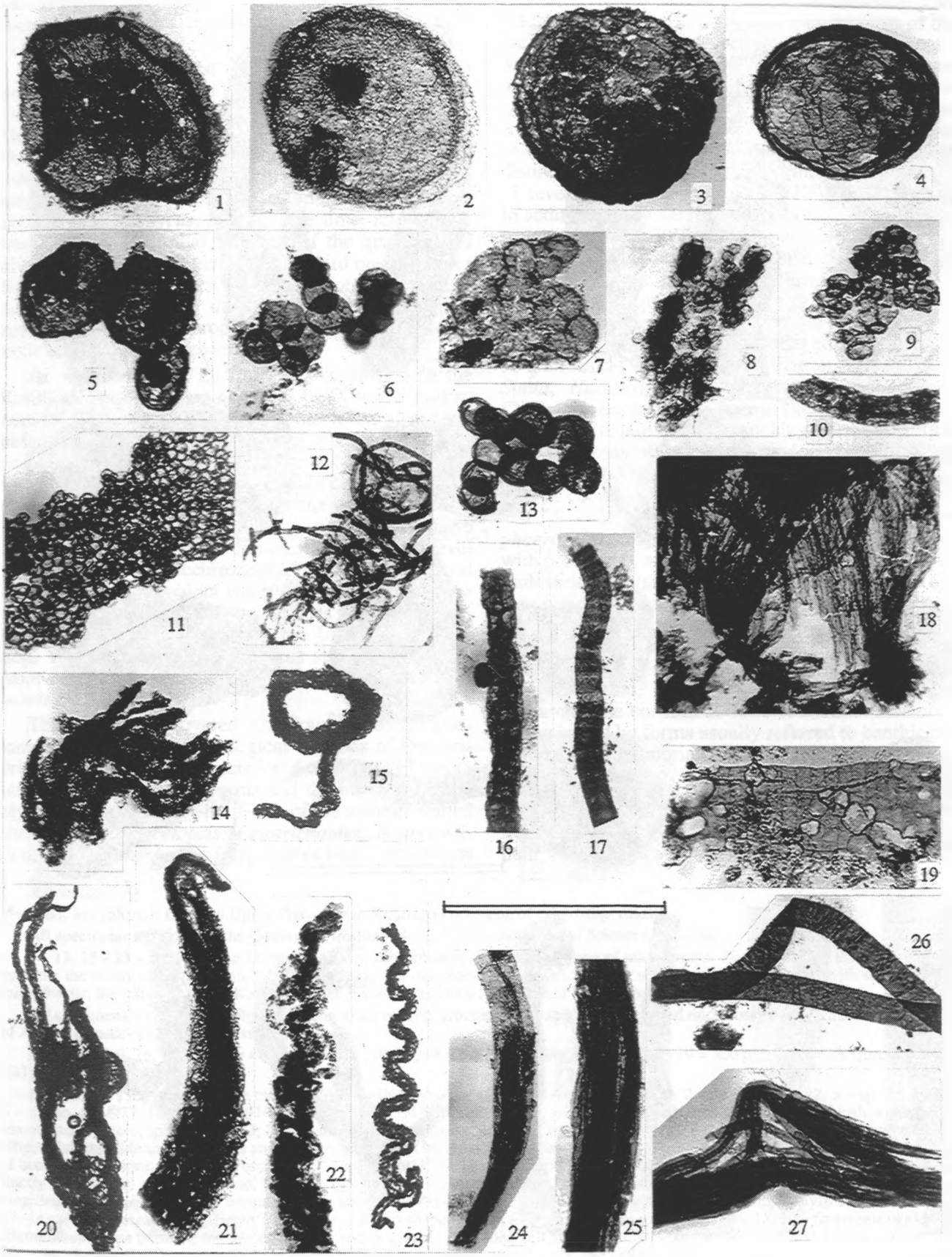

Plate II.

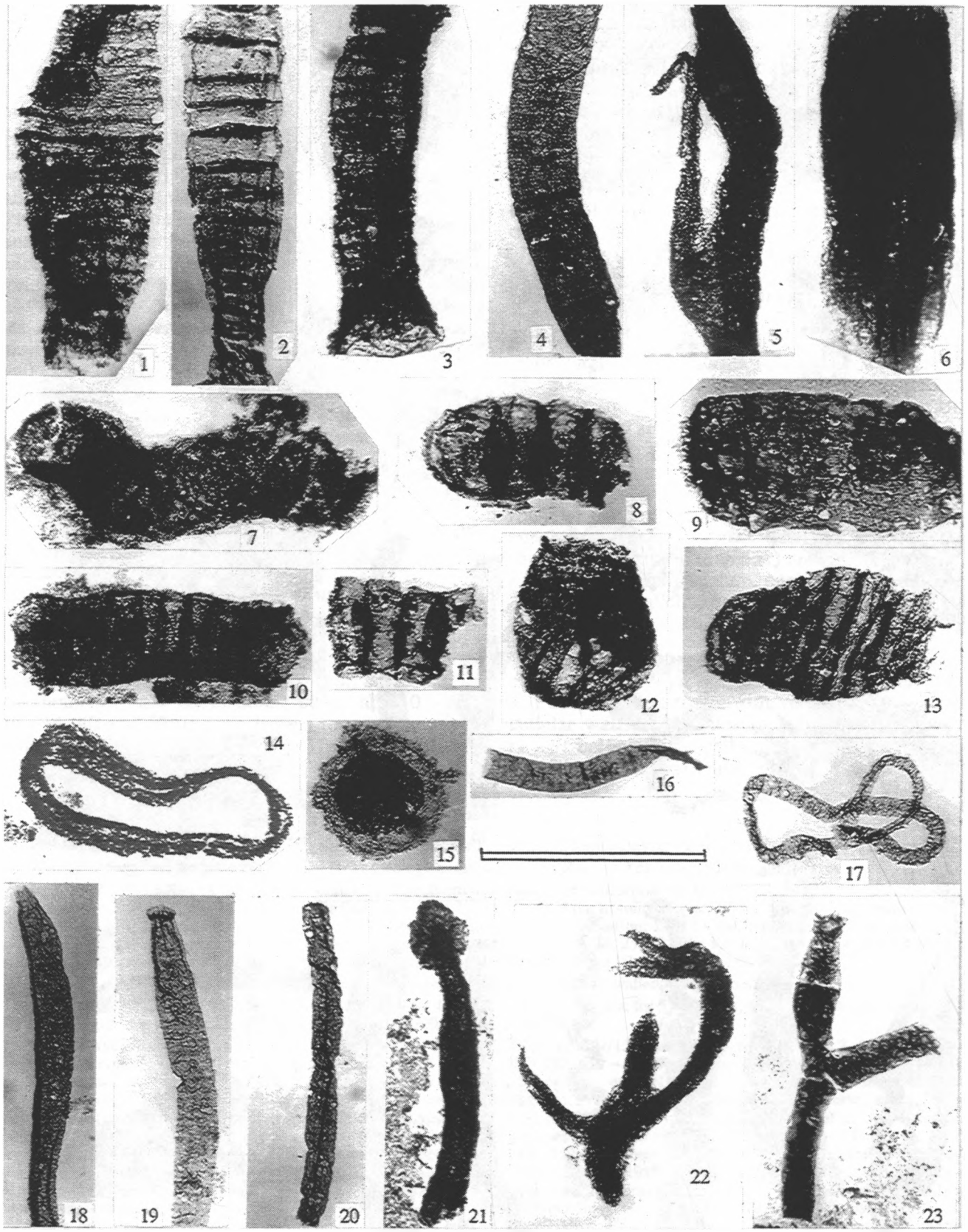

Plate III.

like vesicles with dark inclusions (the vesicle diameter varies from 160 - 260 μm, the inclusions are 20 - 40 μm in size).

Among the three Ust'-Il'ya microbiotas, the most peculiar is the first one. It is the only biota that revealed smooth-walled and crinkled-costate *Asperatofilum*, *Taenitrichoides*, *Rectia*, and *Plicatidium* forms and also *Polytrichoides* sheaths, large shrublike *Ulophyton* thalluses, peculiar filmy fossils with a rhombic pattern on their surface, and colonial coccoidal *Myxococcoides* and *Coniunctiophycus*. At the same time, all of these three biotas are similar in terms of the quantity and diversity of their acritarchs, but the third one surpasses two others in abundance and dimensions of the thalluses. On the whole, the Ust'-Il'ya biota displays approximately equal proportions of supposed planktonic and benthic representatives.

As we mentioned, all finds of microfossils in the Kotuikan Formation are confined to its lower subformation. The Kotuikan biotas at the levels described before (Veis and Vorob'ieva, 1992) seem to be impoverished versions of the more representative Ust'-Il'ya assemblage. In contrast, the recently discovered biota (130 m above the formation base, the locality near the mouth of the Ulakhan-Tomulakh Creek) proved to have a more diverse taxonomic composition (the sixth level of the microfossil occurrences, Fig. 3). The general composition of this biota is very similar to that of the rich Ust'-Il'ya biotas, but also includes some characteristic morphotypes unknown in the Ust'-Il'ya Formation. We shall see below that this similarity resulted from comparable environmental settings of sediments bearing these biotas.

The recently discovered Kotuikan microbiota contains all the main morphological varieties of Precambrian organic-walled microfossils. Some of them (especially filamentous forms and thalluses) are notably large. For example, the width of the smooth-walled sheaths of *Eomycetopsis*, *Leiotrichoides*, *Asperatofilum*, and *Taenitrichoides* forms varies from 2 to 100 μm and even more. The width of trichomelike forms reaches 240 - 260 μm (morphologically, most of them are similar to the previously described genera of hormogonial cyanobacteria *Botuobia* and *Oscillatoriopsis*). These forms associate with numerous *Polytrichoides*, spiral-circular *Elatera*, thinning towards their ends *Siphonophycus*, and the specific structures of two varieties – the ribbonlike forms (50 μm in width) with distinct longitudinal pattern, and the fossils consisting of several ring-shaped elements 120 μm in diameter. In addition to filamentous organic-walled microfossils, which comprise 30% of the sample selection under study, the lower Kotuikan Subformation yielded numerous acritarchs and ribbonlike thalluses. Both groups are almost equal in quantity, comprising 30 - 35% of all the forms found there. The acritarchs vary in size from 10 to 1400 μm (the dimensional peaks correspond to 150, 300, and 500 - 700 μm) being represented by *Chuaria*, *Simia*, *Nucellosphaeridium*, *Pterospermopsimorpha*, *Cucumiforma*, and other forms. The dark, structureless thalluses (including the branching forms) consist of several dozens of specimens, and vary in width from 16 - 18 to 160 μm. Here we also found extended colonies of coccoidal *Ostiana* (with a cell diameter of 10 and 20 - 25 μm), fungilike fossils (the *Germinosphaera*, *Caudosphaera* and/or *Vendomyces* vesicles with processes), and branching cylindrical sheaths with multiserial chainlike aggregates of coccoidal cells, resembling in structure some representatives of recent stigonematacean cyanobacteria. In addition, we distinguished several broad filaments (up to 80 μm wide) with attached spherical vesicles, probably belonging to *Vanavarataenia*. On the whole, despite its abundant acritarchs, the considered lower Kotuikan assemblage is dominated by forms usually referred to benthic components of Precambrian biotas.

As stated above, using the Turukhansk Riphean organic-walled microfossils as an example, we identified a series of their environmental groupings. Some of their analogues can be unambiguously recognized in

Plate III. Microfossils from the Upper Precambrian Kotuikan Formation of the Anabar massif.

All specimens are stored at the Geological Institute of the Russian Academy of Sciences, Moscow.

1 - 13, 15 - 23 – the Kotuikan Formation, lower subformation, a new occurrence of microfossils; the left bank of the Kotuikan River, opposite the mouth of the Ulakhan-Tongulakh Creek; 14 – the Kotuikan Formation, lower subformation, a new sight with solitary finds of microfossils; the left bank of the Kotuikan River, 6 km downstream the mouth of the Ulakhan-Tongulakh Creek.

All specimens, except that in Fig. 14, belong to the optimal grouping of Anabar organic-walled microfossils; 14 – ? transitional grouping of Anabar organic-walled microfossils.

Bar for 1, 6, 8, 16, 17, 20, 22 = 400 μm; for 2, 5, 10, 11, 13, 15 = 330 μm; for 3, 4, 7, 12, 18, 19 = 300 μm; for 9 = 120 μm; for 14, 23 = 100 μm; for 21 = 1000 μm.

1 - 3, 8, 10, 11, 13 – very large trichomelike fossils of unclear affinity: 1 – sp. no. 591/970, 2 – sp. no. 591/972, 3 – sp. no. 591/971, 8 – sp. no. 591/977, 10 – sp. no. 591/979, 11 – sp. no. 591/980, and 13 – sp. no. 591/982; 4 – a broad filamentous sheath with distinctly transverse structure; sp. no. 591/973; 5, 22 – large branched thalluses: 5 – sp. no. 591/974 and 22 – sp. no. 591/991; 6 – a large elongated-ellipsoidal form thinning to its ends and displaying a distinct longitudinal structure; sp. no. 591/975; 7, 12 – *Elatera binata* Herm.; fragments of broad tubular springlike microfossils and 7 – sp. no. 591/676; 12 – sp. no. 591/981; 9 – a broad elongated (? ribbonlike) form with distinctly longitudinal structure; sp. no. 591/977; 14 – a large fossil composed by two ring-shaped elements; sp. no. 891/383; 15 – *Pterospermopsimorpha pileiformis* Tim. emend. Mikh.; sp. no. 591/984; 16 – *Palaeolyngbya sphaerocephala* Herm. et Pylina; sp. no. 591/985; 17 – *Asperatofilum experatus* (Herm.); fragment of a loop-shaped curved filamentous form; sp. no. 591/986; 18, 19 – fragments of elongated filamentous forms thinning to their ends (19 – with a specific expanded cell at the end): 18 – sp. no. 591/978 and 19 – sp. no. 591/988; 20 – an elongated filamentous form without thinning at rounded, closed ends; the central part of the form has a narrow dark zone; sp. no. 591/989; 21 – a germinated reproductive structure of algae or of fungi; sp. no. 591/790; 23 – a branching cylindrical sheath with a multiserial chainlike aggregate of coccoidal cells; sp. no. 591/918.

the succession of the Ust'-Il'ya and Kotuikan microbiotas. Judging by taxonomic composition, dimensions of organic-walled microfossils, and their characteristic morphotypes, the three successive Ust'-Il'ya biotas can be related to the "optimal" grouping, while the assemblages of small acritarchs and filamentous sheaths near the base of this formation can be considered representatives of the shallow or transitional groupings. Such an interpretation is completely consistent with the environmental features of the enclosing beds, which recorded settings of the initial stage in the rapidly developing transgression. Among the deep-water deposits, the three-meter thick member of shales near the base of the upper Ust'-Il'ya Subformation yielded only abundant films of brown, decomposed organic matter.

The newly discovered lower Kotuikan biota undoubtedly belongs to the "optimal" grouping, as well. Like the similar grouping of the Ust'-Il'ya organic-walled microfossils, it contains such indicative morphotypes as very large acritarchs, colonial coccoidal forms, thalluses, and fungilike fossils. Against this background, there are some gigantic trichomelike bodies inherent only to this biota. According to their dimensions, they exceed all previously known Riphean microfossils of similar structure (except *Segmentothallus* from the lower Lakhanda Subformation). Judging by abundant thalluses and fungilike forms, the lower Kotuikan biota had a relatively deeper-water origin in the zone of the "optimal" grouping distribution.

As for the three lower Kotuikan biotas described previously, only one of them (stratigraphically the lowest) may be, with certain reservations, related to the "optimal" grouping. This biota is considerably inferior in the diversity and dimensions of its forms than the one discussed above, but the two are similar in terms of their characteristic morphotypes. Their differences may be attributed to the slightly shallower setting of the less diversified biota, as is evident from sedimentological data (? the proximal part of the zone of the "optimal" grouping distribution). The other lower Kotuikan organic-walled microfossils probably represent the shallow-water groupings of such biotas extremely rarely recorded in carbonate platform settings. Their taxonomic diversity is sharply reduced (rare small acritarchs, solitary fragments of narrow sheaths, and thalluses), and the fossils occur in thin clayey-carbonate interbeds in essentially carbonate deposits. The low diversity of these biotas may be explained by the unfavorable environmental conditions and by the well-known adverse influence of the carbonate medium on the preservation of organic-walled microfossils (Veis and Semikhatov, 1989; Veis and Vorob'ieva, 1992).

Thus, the Anabar organic-walled microbiotas characterize a sufficiently wide spectrum of paleosettings. The general pattern of their environmental distribution is, despite some specific features, in general identical to that for organic-walled microfossils from the Turukhansk Uplift.

In addition to the organic-walled microfossils considered above, the Kotuikan Formation yielded a variety of silicified microorganisms from both its lower and upper subformations (Golovenok and Belova, 1984, 1986; Yakshin, 1991; Sergeev, 1993, and others). In total, the Kotuikan assemblage of mineralized forms includes more than 14 genera and 13 species of microfossils. The predominant among them are long-ranged Precambrian taxa, such as colonial coccoidal *Eontophysalis*, *Myxococcoides*, *Gloeodiniopsis* and filamentous *Eomycetopsis*, *Siphonophycus*, *Oscillatoriopsis*, and *Paleolyngbya* in association with the rare spheroidal *Eosphaeronostoc*, and also especially large *Eosynechococus* and the mat-forming filamentous *Circumvaginalis* known only in the Anabar section. Most investigators of the mineralized Kotuikan biotas emphasized the abundance and very large dimensions of the remnants found there; for instance, the trichomelike forms up to 85 μm in width, and vesicles exceeding 400 μm in diameter. All these large forms originate from the upper Kotuikan Subformation (Yakshin, 1991; Sergeev, 1993). Paleoenvironmental interpretation of the silicified Kotuikan biota, based upon the actualism principles, led Sergeev (1993) to the conclusion that this assemblage was formed within a wide spectrum of settings rapidly changing with time from the subcontinental (amphibial landscapes) to the open-marine environments. The appearance of diverse assemblages with the largest and most complex morphs was related by this author (Sergeev, 1993, p. 48) to the short periods of unusual extremely favorable conditions.

Our data demonstrate that the early diagenetic fossiliferous cherty concretions from the Kotuikan Formation are mostly confined to sediments formed in the shallowest settings. Such settings dominated widely during the late Kotuikan time, and sporadically occurred in the early Kotuikan time (Fig. 3). At first sight, this conclusion does not correspond well to the fact that large and/or complex silicified microfossils actually belong to biotas from the upper Kotuikan Subformation. However, a more detailed analysis of paleoenvironmental confinement of such morphotypes resolves this contradiction: all of these microfossils originate solely from relatively deeper-water sediments of the upper subtidal zone (for example, from the columnar-stratiform stromatolites of *Omachtenia*), which were formed during short periods of elevated sea level. There is a characteristic peculiarity emphasizing the sharply variable pattern of the late Kotuikan sedimentation. All the remarkable morphotypes described above occur in some samples only several millimeters apart from palmelloidal colonies of *Eontophysalis*, scoured surfaces, and some other positive signs of the extremely shallow-water settings.

Thus, the distribution of the Anabar silicified biotas also shows a certain spatial ordering, which corresponds to the above considered paleoecological structure of Riphean biotas. Moreover, the silicified biotas have their own place in this structure. The taxonomically poor assemblages dominated by benthic mat-

forming *Eontophysalis* and accompanying forms are enclosed in very shallow-water deposits, and can certainly be related to the shallow-water grouping. At the same time, the more diverse communities, containing large and/or complex morphotypes and represented by approximately equal quantities of coccoidal and filamentous, mainly allochtonous forms, are related to other, more deep-water groupings, up to the "optimal" one.

Following from the above, the two regarded regions of central and north Siberia, namely the Turukhansk and the Anabar areas, reveal the uniform main features of lateral distribution of organic-walled and silicified Riphean microfossils.

The Uchur-Maya Region

The Riphean succession of this region is among the world's stratigraphically most complete and microphytologically best studied. Unfortunately, it lacks detailed sedimentological analysis. Nevertheless, we shall summarize previously published data on microfossil distribution in the Omakhta and Totta formations of this region (Veis and Semikhatov, 1989; Veis and Vorob'ieva, 1992), and consider these data within the context of the presented model of the environmental distribution of Riphean microfossils.

The Omakhta microfossils are of particular interest because they permit us to evaluate the applicability of the above model to relatively shallow-water basins with very variable settings and siliciclastic-carbonate sedimentation. The Omakhta deposits are noteworthy in their small cycle units (Semikhatov and Serebryakov, 1983). Most of the Omakhta cycles are 3 - 5 m thick, and generally correspond to transgressive sequences. Typically, the lower cycle parts contain siliciclastic rocks, while the upper parts consist of dolomites (sandy, detrital, oncolitic, and common stromatolitic varieties). The cycles show significant lateral variations that suggest a highly mixed facies distribution and frequently alternating shallow basin environments.

All the Omakhta microfossils were found in one section (about 450 m thick), situated in the southern part of the Uchur depression along the Idyum River (Veis and Semikhatov, 1989). The microfossils were discovered at six successive levels, confined to three consecutive units (130 - 140, 160 - 180, and 48 - 50 m thick) differing in cycle compositions. The cycles of the lower unit contain a relatively high proportion of sandy members with many desiccation cracks, ripple-marks, and polymodal coarse cross lamination. Their carbonate members are dominated by clastic dolomites with small scour surfaces, lenses of endoclastites, and salt crystal moulds. In the second, and especially in the third units, the sandstones and marks of extremely shallow-water setting are less abundant; the lower cycle parts are richer in shales whereas their upper parts contain glauconite globules, and traces of subaerial exposures seem to disappear. All these features suggest that the deposits under consideration were formed in a wide range of settings characteristic of tidal flats, ephemeral shallow lagoons (the lower and the middle units), and protected settings of the upper subtidal and the lower intertidal zones (the upper unit).

The six Omakhta microfossil assemblages embrace 32 species and 25 genera, representing acritarchs, colonial coccoidal and filamentous forms, associated with intricate morphs (Veis and Semikhatov, 1989). These biotas display directional changes of their taxonomic composition and overall dimensions of morphotypes related to environmental diversity in the Omakhta paleobasin. Omitting details, we may say that the three first biotas, confined to the lower unit of the Omakhta Formation, are of a very low taxonomic diversity. Only several varieties of acritarchs (*Leiosphaeridia*, *Protosphaeridium*, and *Nucellosphaeri-dium*) were found here in association with the *Cyanonema* and *Oscillatoriopsis* trichomes; the former do not exceed 55 μm in size, and the latter are 8 - 10 μm wide. In addition to these, taxa the fourth biota found in the middle unit also contains the ellipsoid-shaped acritarchs *Konderia*, marginate *Simia*, laminated-globular *Polycavita* and *Stratimorphis*, colonial coccoidal *Coniunctiophycus*, filamentous *Siphonophycus*, and processes-bearing vesicles of *Germinosphaera*. Compared to the previous biotas, the overall dimensions of all morphotypes in the fourth microbiota are larger (for acritarchs, which are up to 130 μm, and for filamentous forms – up to 18 - 19 μm). The fifth and sixth Omakhta biotas, confined to the upper, relatively deep-water unit, are distinguished by a much higher taxonomic diversity and by significantly larger microfossil dimensions. These biotas contain almost all the above varieties of organic-walled microfossils and also such forms as acritarchs *Pterospermopsimorpha*, tabular *Satka*, double-layer pseudo-spinous *Eomarginata*, colonial coccoidal *Palaeopleurocapsa*, filamentous *Filiconstrictosus*, *Leiotrichoides*, and some others (here, acritarchs are 150 μm and more in size, the filamentous forms, 20 - 22 μm).

All the compositional and environmental data on six Omakhta biotas allows us to attribute the first three of them to the shallow-water ecological grouping, and the fourth, taking into account its position in the section, to the transitional one. The richest Okhmata microbiotas may be, with certain reservations, related to the "optimal" grouping, identified in the Bezymyannyi and Ust'-Il'ya–Kotuikan formations. Such a definition is favored not so much by the compositional similarity of these biotas (which is undoubtful), but rather by their common position on the bathymetric profile and their similar saturation in specific morphotypes. At the same time, the two richest Omakhta biotas considerably differ from typical "optimal" groupings in other investigated regions, where they are very uniform and persistent in composition. The differences are vividly manifested in the incomparable dimensions of the forms found in these biotas (acritarchs and the filamentous forms from Turukhansk and Anabar localities are

nearly an order of magnitude larger than the respective Omakhta fossils).

These differences may be attributed either to the specific features of the Omakhta basin, which was characterized by its carbonate-siliciclastic sedimentation and more widespread shallow-water settings than the Bezymyannyi, Ust'-Il'ya and early Kotuikan basins, or, as one of the authors believes, to the different age of the Omakhta Formation relative to the Ust'-Il'ya-Kotuikan and Bezymyannyi deposits. Accepting the second opinion, we may suggest that the above dissimilarities between the Omakhta and other "optimal" groupings under discussion reflect the evolution of the environment-dependent variations in the composition of the Riphean microbiotas.

As for the 12 microbiotas of the Totta Formation (Veis and Vorob'ieva, 1992), we should mention that they were distinguished in the stratigraphical analogs of the Bezymyannyi Formation and seem to be similar to its assemblages. The Totta organic-walled microfossils were discovered in a vast area at 12 stratigraphical levels. Five of these levels belong to the lower Konder Subformation (the Bol'shoi Aim River, the northern slope of the Omnya uplift), and seven other levels – to the upper Omnya Subformation (the Omnya, Aim and Maya rivers, the Maya depression).

The Konder deposits (120 m thick) form a distinct transgressive-regressive rhythm. At the base and top, the subformation is represented by thick members of light-grey or grey, medium- to coarse-grained, usually cross-laminated, glauconite-bearing sandstones, and subordinate cross- and horizontally-laminated siltstones. The middle member of the subformation contains greenish and reddish units of frequently alternating siltstones and shales with fine horizontal or gentle cross lamination, which enclose scattered globules of glauconite, siderite concretions, and scour channels (of the gutter casts type) in some layers. The Omnya Formation (400 - 600 m thick) shows a monotonous alternation of small packages of grey fine-grained cross-laminated sandstones with frequent siltstone interbeds and thick units of greenish to brownish, fine-laminated siltstones and shales.

The most representative and diverse biotas of the Totta organic-walled microfossils were found in the middle parts of the Konder and the Omnya subformations (the third and the seventh biotas in the entire Totta succession). They were confined to a certain zone of the paleobasin with medium depths and relatively quiet hydrodynamic conditions. Both biotas include: particularly large acritarchs *Chuaria* (with diameter up to 700 μm and even 1000 - 1100 μm); extended thalluses of coccoidal *Ostiana* (the cell diameter from 10 to 20 μm and more); broad (up to 40 - 60 μm) smooth-walled and/or crinkled-costate (up to 55 μm in width) sheaths of *Taenitrichoides*, *Rectia*, and *Rugosoopsis*; multifilamental *Polytrichoides*; cylindrical or thinning filamentous *Siphonophycus*; and other taxa known from the "optimal" groupings of the Turukhansk and Anabar regions. Among the striking features of the Totta biotas we must mention extremely abundant specific multiserial colonies of pleurocapsular cyanobacteria *Sphaerocongregus* of several dimensional classes. All these taxonomic and environmental features of the richest Totta microbiotas permit us to relate them to the "optimal" ecological grouping.

A closer similarity is observed between shallow-water groupings of the Uchur-Maya and other regions. In the Totta Formation, such groupings include the first, second, and sixth biotas discovered in the siltstone-sandstone units with structures typical of extremely shallow-water sediments. In their generic and species composition (common species of genera *Leiotrichoides*, *Myxococcoides*, and *Leiosphaeridia*) and microfossil dimensions, they are very similar to biotas characteristic of the shallow-water Bezymyannyi deposits except for the presence of solitary ellipsoidal *Eosynechococcus* and fragments of chainlike *Sphaerocongregus*, known from the Totta Formation.

The fourth and fifth Totta biotas were found very near to red shale members of the Konder Subformation. From the standpoint under consideration, they may belong to the distal part of the "optimal" zone. Their composition is dominated by: abundant acritarchs with a diameter under 500 μm; by sheaths up to 40 μm wide, and by numerous chainlike *Sphaerocongregus*, extended *Ostania*, multifilamental *Polytrichoides*, and branching *Ulophyton* thalluses (the "bush" is over 700 μm high).

The deepest-water microfossil groupings in the Uchur-Maya region are most likely represented by the eighth - tenth Omnya biotas, confined to monotonous shale-siltstone members with faint lamination. In a mass of multiple films of decomposed brownish organic matter there are rare acritarchs of 70 - 100 μm in diameter, scattered sheaths of 2 - 20 μm in width, and fragments of coccoidal *Myxococcoides* with cell diameter of 20 - 30 μm. The impoverished composition of these biotas can be considered as an indication of their affinity with shallow-water groupings. But such a suggestion is inconsistent with the character of the enclosing beds and, in particular, with abundant organic films in them. Taking this into account, these three biotas can be related to the deep-water grouping of organic-walled microfossils, which was earlier identified among the Linok and Ust'-Il'ya materials.

Thus, we may state that general features of the taxonomic composition of the Totta organic-walled microfossils and the environmental confinement of their assemblages correspond to those described for the Turukhansk and Anabar sections of the Riphean in Siberia.

THE POSITION OF THE ORGANIC-WALLED MICROFOSSILS ASSOCIATIONS IN THE SIBERIAN RIPHEAN BASINS

The Riphean deposits of the Turukhansk, Anabar, and Uchur-Maya regions generally represent very similar, large transgressive sequences that reflect prograd-

ing marine transgressions and related changes in the rock composition, i.e., a transition from siliciclastic, mainly clayey sedimentation to the carbonate deposition. Such a transition was most likely related to the remoting provenances and pertinent shortage in the continental siliciclastic material supply. Simultaneously, the main paleoecological and environmental parameter, the paleodepth also changed, though not gradually. In all of the above reconstructed examples, the depth maximum is manifested in the transition from siliciclastic to carbonate types of sedimentation (Fig. 5). This peculiarity seems to be related not only to the reduced input of siliciclastic material into the basin, but also to some delay in the generation of intrabasinal carbonates. When initiated, the carbonate sedimentation became more and more important, and eventually resulted in the appearance of extremely shallow tidal-flat settings (an amphibial landscape) with extreme living conditions for microorganisms. Most likely, the appearance of such settings was also a result of a progressively reducing subsidence rate.

Based on the discussed changes in the paleoenvironments and style of sedimentation, we may identify four types of paleobasins:[2] (Fig. 5). 1. Basins with siliciclastic sedimentation and a wide range of settings existed at the early stage of transgressions. Unlike other basins considered here, these had the largest bottom relief amplitude, and the settings ranged from alluvial plains (the Labaztakh time on the Anabar massif) to relatively deep-water open-marine zones (the northern part of the Turukhansk uplift during the early Bezymyannyi time). 2. Relatively deep open-marine basins with mixed carbonate-clay sedimentation, chiefly below the storm wave base (the Western Anabar area at the late Ust'-Il'ya time and the Turukhansk uplift at the Linok time). 3. Basins of medium depths and mainly carbonate sedimentation (the Western Anabar area at the early Kotuikan time, and the Turukhansk uplift during the early Sukhaya Tunguska time). Most probably, these basins extended over gently sloping carbonate platforms, dominated by an accumulation of clastic carbonates, and shelves with various reef structures and a characteristic variety of facies. 4. Basins of the final regressive stages with extremely shallow settings and purely carbonate sedimentation (the late Kotuikan and the late Sukhaya Tunguska time). A characteristic feature of their deposits is a wide occurrence of early diagenetic chert nodules.

The presented data permits us to make the following main conclusion: The taxonomic diversity, and to a certain degree the abundance of microfossils in the Riphean basins of Siberia, were strictly dependent on the basin depth and did not reveal a clear correlation with the main topographic features of the basin floor and type of sedimentation.[3]

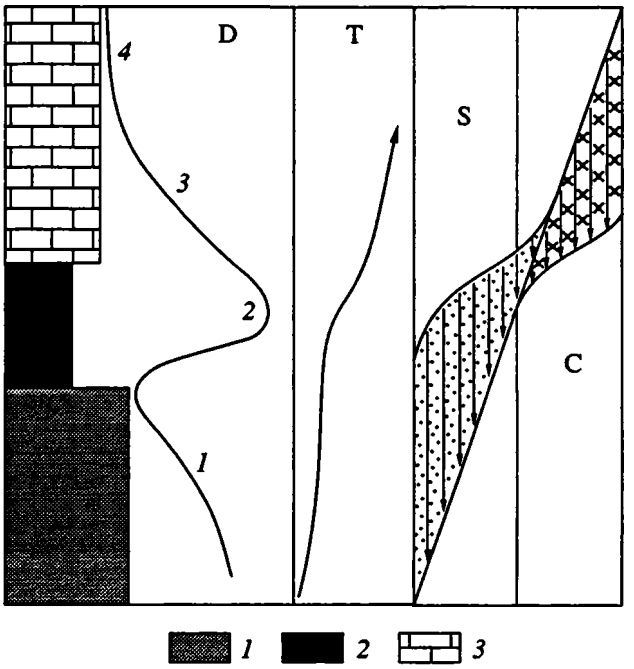

Fig. 5. Generalized structural scheme of a large transgressive cycle in the Siberian platform Riphean and the evolution of the depositional environment during the formation of this cycle.

D – curve of depth variations in the basins. Numbers indicate the basin types (see text). T – curve of the transgression development. The type of sedimentation: S – siliciclastic, C – carbonate. The length of arrows is proportional to the sedimentation rate. Composition of deposits: *1* – clayey-sandy; *2* – silty-clayey; *3* – carbonate (limestones and dolomites).

It is noteworthy that the available materials indicate that the studied Riphean microfossil assemblages reflect their primary distribution in the paleobasins, but not the result of their postmortem transportation and burial. First of all, this conclusion is based on the directly observed environmental confinement of microfossils: in all cases, the most representative "optimal" groupings were found in deposits accumulated within the same narrow depths range (10 - 30 m), regardless of the fact that in various basins this range of depth corresponded to shelves that differed greatly in their relief, i.e., to those of contrast topography (the first Bezymyannyi optimal grouping) and to the nearly flattened platform (the second Bezymyannyi "optimal" grouping). On the other hand, the above conclusion is confirmed by analyses of the taxonomic and ecological compositions of the groupings, among which the "opti-

[2] Here and below, by the term "paleobasin" we mean a part of the paleogeographic media with a definite set of similar paleosettings that existed when large sedimentary units, namely formations or subformations, were accumulated. Such similar paleosettings (the environments of sedimentation) resulted in petrographical and structural affinities of rocks from the respective lithostratigraphical units. In the strict paleogeographic and structural sense, the paleobasins identified here correspond to certain parts of a sedimentary basin.

[3] In this respect, other paleoenvironmental factors could also be of importance, namely the salinity and temperature regime, hydrodynamic activity of the basins, and others. However, neither our, nor published data provide adequate grounds to objectively analyze the role of these factors.

mal" one contains all the most demonstrative morphotypes (both of plankton and of benthos) that are similarly very well preserved and presented in comparable quantities.

The most general characteristics of microfossil distribution may be described as follows: their taxonomic diversity, structural complexity, and dimensions were rapidly increased in passing from the deepest-water to medium-deep parts of the siliciclastic basins. The trend of changes was not linear, since the above parameters reveal a sharp maximum confined to a depth interval of several dozen meters. Towards shallower basin areas with siliciclastic sedimentation, the biotas rapidly lost their taxonomic diversity and peculiar procaryotic benthic assemblages appeared within zones of carbonate sedimentation. Below we consider this distribution model in more detail for each of the inferred basin types.

1. Basins with siliciclastic sedimentation and a wide range of settings. Four paleoenvironmental groupings of organic-walled microfossils, the deep-water, "optimal," transitional, and shallow-water, were distinguished in the deposits of these basins. All of them contain some common or transit taxa of small acritarchs and fragments of colonial coccoidal and filamentous forms (*Leiosphaeridia*, *Myxococcoides*, *Eomycetopsis*, and *Leiotrichoides*). These representatives define the limits of taxonomic diversity in the shallow-water groupings. The diversity and dimensions of microfossils in the transitional groupings are higher, and reach their maximum in the "optimal" groupings. It is only the latter that include especially large acritarchs *Chuaria*, peculiar *Simia* and *Konderia*, vesicles with processes, extended *Ostiana* colonies, chainlike *Polysphaeroides*, multifilamental *Polytrichoides*, gigantic trichomelike bodies, branching *Ulophyton* and *Archaeoclada* thalluses, segmented *Eosolena*, as well as filmy, tubular, ribbed, and many other forms. As it was demonstrated, the taxonomic diversity of microfossils in such basins rapidly decreased towards zones of greater and smaller depths, but in different manners for shallow and deep settings. The deep-water groupings of microfossils include assemblages rich in thalluses and fungilike (hyphomicrobial) fossils, while the transitional and shallow-water groupings are enriched in relatively simple acritarchs of medium dimensions, colonial coccoidal, and filamentous forms. Besides their taxonomic composition, the groupings sharply differ in their ecological populations. The shallow-water and transitional types chiefly include allochthonous (plankton) microbial associations, while the optimal and deep-water groupings are basically benthic.

2. Relatively deep-water open-marine basins with mixed carbonate-clay sedimentation. The available data do not allow us to recognize any essential differences in the paleoenvironments and the microphytological parameters between these and the above deep-water basins with siliciclastic, or slightly carbonate clayey deposits. Indeed, the lower Bezymyannyi clayey deposits (basins of the first type) and the carbonate-clayey Linok deposits (basins of the second type) contain similarly impoverished microfossil assemblages dominated by the same morphotypes, namely acritarchs of medium and large dimensions, thalluses, fungilike forms, and subordinate trichomelike bodies. The peculiar feature of the carbonate-clayey deposits in the basins of the second type is their enrichment in thin films of dispersed organic matter devoid of definite microstructure and any distinguishable microfossils. As was noted, such films seem to be the remains of mats formed by microbial benthic communities. Similar films, referred to as mat remains, were described in similar deposits of the Riphean Newland Formation of the Supergroup Belt, USA (Schieber, 1986, 1990). The clayey-carbonate rocks of this supergroup, similar to the Linok deposits, contain rare filamentous microfossils that, most probably, were mat-forming (Horodyski, 1980). According to Horodyski, the scarcity of these microfossils can be attributed either to their small dimensions or to their low stability under aggressive environmental factors. It seems plausible that some destructors also contributed to impoverishment of those fossils (Oehler, 1977).

3. Basins of medium depths and chiefly carbonate sedimentation. Unfortunately, the bulk of microphytological information on these basins seems to have been lost, due to both a low content of clayey rocks, which favor organic-walled microfossil occurrences, and the adverse effect of the alkaline diagenetic environment upon the microbial organic matter (Veis and Semikhatov, 1989). It is not always possible to compensate such informational losses by studying the mineralized biotas. The latter are rather rare in the examined Precambrian sections, because chert nodules available here are mainly of late diagenetic origin and barren of fossils. In our samples, only two biotas can be assigned to the assemblages of these basin types: the one recently discovered in the lower Kotuikan Subformation (see above), and some other mineralized microfossils present there (Yakshin, 1991; Sergeev, 1993). Both assemblages comprise characteristic morphotypes, including similar acritarchs, broad trichomelike forms, and some other microfossils, all identifiable as "optimal" grouping of microorganisms. The prevalence of benthic forms in these biotas corresponds to such an identification. It is noteworthy that depositional environments within the vast carbonate platform of these basin types suggest rapid changes in the water depth and, accordingly, in the sedimentation regime of the paleosettings. Such changes are reflected by two lower Kotuikan assemblages of mineralized microfossils, which occur near each other and drastically differ in their taxonomic and ecological composition (Sergeev, 1993).

4. Basins with extremely shallow-water settings and purely carbonate sedimentation. All the available information on the microbial inhabitants of these basins is exclusively based on an examination of mineralized microfossils. Their assemblages were studied very thoroughly, an apparent contradiction that can be easily explained. On the one hand, the shallow peritidal parts

of the basins concentrated the bulk of early diagenetic cherty concretions that proved to be excellent sarcophaguses for microbiota. On the other hand, it was these very zones of the Riphean seas that were densely populated by assemblages of mat-forming procaryots following the spread of large masses of stromatolitic carbonates. Today, scientists unanimously estimate the nature and paleogeographical position of these assemblages, considering them the most important component of the extremely shallow-water ecosystems of tidal flats. The fact that ancient tidal flat deposits reach dozens and hundreds of meters in thickness, sometimes traceable in lateral direction for hundreds of kilometers (Grotzinger, 1986, 1989), strongly suggests that such paleolandscapes have no recent analogs, though they resemble modern coastal marine zones in depth range and sedimentation pattern.

Without going into the details of well-known data on the taxonomy and paleoecology of shallow-water silicified microfossils (e.g., Knoll and Golubic, 1979; Knoll *et al.*, 1991, and others) we shall stress only the differences between the shallow-water groupings in zones of carbonate and of siliciclastic sedimentation. First, the organic-walled microfossils in the shallow zones of siliciclastic accumulation were found to be the poorest both in taxonomic diversity (very rare and simple acritarchs, colonial coccoidal and filamentous forms) and abundance. They are represented by allochthonous elements, but at the same time, the shallow-water groupings of carbonate platforms contain benthic microbiotic communities, which are much more diverse and abundant. Such communities comprise clearly distinct taxa of mat-forming microorganisms, and also symbiotic and allochthonous forms. Second, poorest, taxonomically speaking, shallow-water groupings of organic-walled microfossils that inhabited zones of siliciclastic sedimentation do not show any spatial variations in their composition. Similar groupings of carbonate platforms, on the contrary, clearly demonstrate compositional variations related to their position in concrete zones of paleobasins (lagoon, or barrier assemblages, and others; Knoll *et al.*, 1991). However, the shallow-water groupings under comparison shared similarities, which were manifested in their striking "evolutionary" conservatism. Heterochronous assemblages of these groupings that existed hundreds of millions of years earlier or later are surprisingly similar.

Now, ending our analysis of the spatial distribution of the facial-ecological microfossil groupings in the Riphean paleobasins of Siberia, we shall draw some inferences. The available data on biota from the deep-water basins allow us to consider it as a microfossil assemblage chiefly composed of benthic microorganisms with a large biomass and low taxonomic diversity. As inherent elements of such assemblages, there were reducing organisms (probably heterotrophic bacteria and fungi or hyphomicrobes). Occurring below the permeability zone for the most active part of the light spectrum, these assemblages were inevitably enriched in microorganisms that used sources of energy other than light. This can be confirmed by the deep-water biota from the Lower Riphean Barney Creek Formation, Northern Australia, which is saturated with specific bacterial remnants (Oehler, 1977).

As for the "optimal" (medium-deep) microfossil grouping, the available data did not allow us to outline any actual factors responsible for its formation and localization. Moreover, there are grounds to suggest that such "special" factors never existed at all, and that this grouping was distinct from others owing to some non-trivial properties of the Precambrian biota. In this connection, two essential peculiarities of medium-deep zones should be emphasized. First, they were characterized by the fading oscillations of the most significant abiotic factors (hydrodynamic activity, temperature, illumination, and others). Second, these zones were completely dominated by representatives of only one link of the trophic system, namely phototrophs, which produce organic matter, while active or passive consumers of the latter are actually unknown here. This asymmetry of most Precambrian assemblages of organic-walled microfossils repeatedly drew the attention of researchers (e.g., Zavarzin, 1984; Chebanov, 1987) and seems to be a typical feature of these assemblages. All members of phototroph assemblages obviously depend on decreasing light intensity. It seems that the last factor was the main obstacle that prevented these organisms from expanding into deeper zones of the paleobasin. In the opposite direction, towards shallower zones, the set of limiting factors was undoubtedly much wider, although the leading factor seems to be the rapidly growing hydrodynamic activity of the basin environment.

Thus, the niche, occupied by the medium-deep groupings of microorganisms, was an optimal ecological compromise between the most important biological properties of phototrophic microorganisms and abiotic components of the paleoenvironments. This determined the main features of the "optimal" groupings, namely their pronounced polytaxonic composition (at the highest level), the unique character of the most characteristic morphotypes, and the large dimensions of microorganisms, which usually are free of protective structures (except the sheaths of hormogonic cyanobacteria). All these facts strongly suggest that the style of behavior in such groupings was predominantly "individual" (or of organismal type), in comparison to the "K-strategy of existence" (Starobogatov, 1985, and others).

Unlike the "optimal," the shallow-water groupings of organic-walled microfossils from basins of siliciclastic sedimentation were confined to the zones much more affected by physico-chemical environmental factors. Nevertheless, these groupings were very rich in phototrophic organisms. In fact, they were a kind of impoverished "optimal" groupings dominated by morphologically simpler but physiologically more resistant components of biotas.

The conditions in the shallow- and extremely shallow-water zones of carbonate platforms were different.

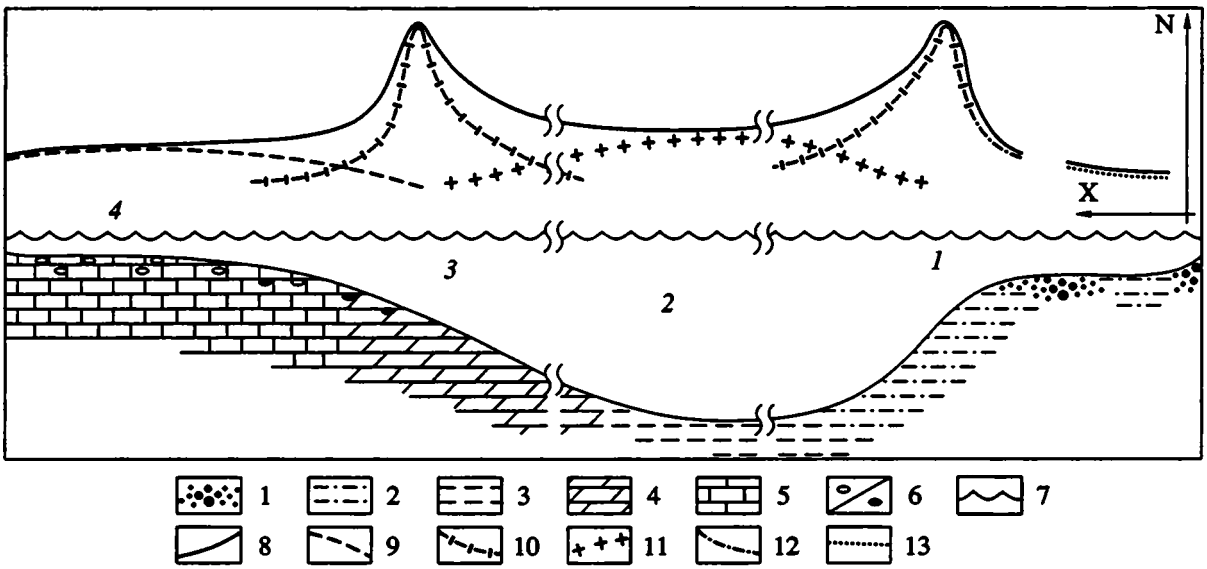

Fig. 6. Distribution of the main paleoenvironmental groupings of microfossils (N) along the profile (X) of various basin types. 1 - 5 – composition of deposits: 1 – sandy-gravelly; 2 – sandy-clayey; 3 – clayey-silty and clayey; 4 – clayey-carbonate; 5 – carbonate; 6 – early and late diagenetic chert nodules; 7 – sea level; 8 - 13 – curves of taxonomic diversity; 8 – the general curve; 9 - 13 – curves for particular paleoenvironmental groupings of microfossils: 9 - shallow-water grouping of carbonate platforms; 10 – "optimal" (medium deep) grouping; 11 – deep-water grouping; 12 – transitional grouping; 13 – shallow-water grouping of basins with siliciclastic sedimentation. Numbers *1 - 4* indicate the basin types (see text).

These zones had extremely unstable physicochemical parameters such as rapid changes in water salinity, temperature, sedimentation rates, gas composition, and, what is especially important, they were periodically subjected to subaerial exposures. The highly rigid abiotic conditions of such paleobasins are reflected in the characteristics of their microfossil assemblages. In response to the aggressive and nonselective influence of the environmental factors, the microorganisms in those basins formed the specific community structure, namely extended, intricately stratified mats (Knoll and Golubic, 1979). Their components of photosynthesis were peculiar in their comparatively low taxonomic diversity, compensated by an immense number of individual specimens organized into colonies and by special protective structures (sheaths, mucilage, and organic matrix). Taking into account the high growth rates typical of such coenoses, as well as their capacity to repair rapidly the damaged structure (Kondrat'ieva, 1975), we can easily understand their high resistance to unfavorable environmental factors (we are probably dealing with the so-called "r-strategy of existence").

The microphytological characteristics of the transitional zones between different facial-ecological groupings of microfossils can be better understood in the case of organic-walled microfossils from basins of siliciclastic sedimentation. The above data on the taxonomic composition of the "optimal" and transitional groupings show that these two biotas gradually replaced one another in the zone of their contact. The higher amount of presumably deep-water morphotypes in the distal parts of the medium-deep paleobasins can be considered proof of the gradual transition between the "optimal" and deep-water groupings. It is much more difficult to understand the spatial interrelations between the principally different "optimal" and shallow-water groupings of carbonate platforms. Biological peculiarities of the shallow-water groupings show that they could have penetrated into much deeper zones, and in fact did so, as is evident from observation data. However, the shallow-water microbiota groupings never seem to have yielded vast benthic coenoses in these zones, and their invasions were of a mosaic (or island) manner. In any case, one should admit that shallow-water and "optimal" groupings from carbonate platforms may have overlapped one another in parts along the bathymetric profile. The transitional grouping, very rarely identified in carbonate deposits, unlike their counterparts in siliciclastic sediments, does not reveal any clear feature distinguishing them from the shallow and "optimal" groupings, but is always a highly reduced and impoverished version of the latter.

CONCLUSION

The main result of this study is the identification of a certain spatial arrangement (structure) in the distribution of microorganisms throughout the Riphean paleobasins. This arrangement resulted in the isolation of some environmental groupings of mineralized and organic-walled microfossils, each having its own taxonomic composition, specific set of indicative morphotypes, the peculiar combination of planktonic and benthic forms, and particular position in the paleobasins of certain types (Fig. 6). The position was defined mainly by such paleoenvironmental factors as the basin

depth and, indirectly, the hydrodynamic activity and light intensity within the setting.

The results obtained in this study naturally continue some earlier inquiries into the environmental distribution of ancient organisms (e.g., Richardson, 1984). Applied to the Precambrian, the model of lateral distribution of microfossils suggested by Knoll (1982, 1984; Knoll *et al.*, 1991) is generally accepted today. However, as one can see from data presented, this model, in fact, reveals only the principal differences between the shallow-water and "optimal" groupings of microfossils from basins of certain types (of the first and third types in our classification), disregarding the characteristic features of all other groupings of microfossils from these and other basins. One of the important consequences of Knoll's model is that the evolutionary transformations in the two above groupings differ in their rates, and, hence, their biostratigraphical potentials are incomparable. Such potential is reported to be the lowest in the shallow-water benthic group of procariots from carbonate platforms, and the highest in planktonic, essentially eucaryotic assemblages of the open-marine basins. Our data permit us to consider this problem from a slightly different viewpoint.

The comparison of the shallow-water and "optimal" groupings of microfossils allows us to draw two important conclusions: 1. Both groupings are, in fact, completely dominated by remnants of phototrophic microorganisms and have no significant traces of heterotrophs that could have selectively destroyed individual representatives of these groupings. Hence, the competition between members for the environmental resources could only be passive (indirect), and the elimination of the present forms could only have happened by accident. 2. Both of the discussed groupings were confined to sufficiently narrow paleoenvironmental zones with their specific habitat conditions. Therefore, it is possible to suggest that both groupings actually happened to be "closed" in their peculiar evolutionary traps, the "optimal" grouping being particularly well trapped. In such cases, recent and fossil organisms show their evolutionary potentials to be stabilizing, their capacity to plastic transformations to be decreasing, the rate of evolutionary changes to be highly retarded, and the populations becoming strictly specialized (Starobogatov, 1985).

Assessing "optimal" groupings of microfossils that are usually regarded as demonstrating the main evolutionary changes in the Precambrian biosphere in this light, it seems important to emphasize the following: Despite the biological diversity of these groupings, they happened to be in an evolutionary dead end. The appearance of the very first more active and less specialized organisms (and especially animals) violated the unique individuality of this particular biotope. Thus, its biota was doomed to vanish.

The slow appearance of morphological innovations among the representatives of the "optimal" groupings, and the almost complete absence of such innovations among the members of the shallow-water groupings are indisputable facts observed in their planktonic and in benthic components. Suffice it say that the entire Riphean history of paleobasins in central Siberia and the South Urals reveals only two or three stratigraphical levels showing such microbiotic transformations that can be considered evolutionary (*Stratotip* ..., 1982; Veis, 1986, 1988; Pyatiletov, 1988; and others). These transformations are fixed in the "optimal" environmental groupings of microfossils. In this context, it is noteworthy that in light of modern ideas on macroevolution, the most favorable biotope for the existence of evolutionary plastic forms is located in the boundary areas of different ecological zones. In our case, the proximal limits of "optimal" zones can be considered as such boundary areas. It was these areas that could be inhabitated by organisms (both autotrophic and heterotrophic) that were not restricted in their evolutionary potentials and could rapidly progress. These very organisms seem to have provided the biological substrate for the wide diversification of the animal kingdom in the later time. The search for remnants of such hypothetical organisms should be regarded as one of the most important objectives of the Precambrian microphytology.

In our discussion of the environmental groupings of Riphean microfossils, we focused our attention on their lateral distribution in the Late Proterozoic paleobasins and did not consider the chronological aspect of their distribution. Meanwhile, a broader comparative analysis of the taxonomic composition and paleoecological confinement of successive Precambrian microbiotas shows that trends in their spatial distribution undoubtedly underwent progressive changes through time. This topic will be analyzed in more detail elsewhere. At the moment, we shall only emphasize that the discussed, rather complicated arrangement of environmental distribution of certain biotas was established only in the Riphean time. It existed about 1100 Ma ago (the Bezymyannyi and Totta formations), but additional studies are required to verify the time of its appearance.

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Evidence of Late Permian Glaciation in the Kolyma River Basin: A Repercussion of the Gondwana Glaciations in Northeast Asia?

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Abstract – The Upper Permian diamicts of the Atka Formation are widespread in the upper reaches of the Kolyma River. They originated not only from seasonal marine ice rafting, as was earlier thought, but were in part also formed by icebergs and possibly shelf glaciers. This suggestion is supported by the complex diamict sequences, the presence of pebbles and boulders with occasional traces of long-term glacier abrasion, the irregular and generally poor roundness of the rock fragments, and the presence of some aggregates resembling till pellets. The diamicts can be termed distal glacial marine sediments, although the presence of proximal glacial marine units cannot be ruled out. In general, the Atka diamicts represent seasonal ice marine and glacial marine sediments, partly reworked by submarine slumps, debris flows, and turbidity currents. The glacial marine sediments found in the Atka Formation are the first find of glacial deposits in the Late Paleozoic Northern Hemisphere, which was confirmed by reliable diagnostic indicators. This finding suggests that the Late Paleozoic climatic asymmetry, relative to the equator, was somewhat less than it previously seemed, though the problem of the climatic asymmetry itself remains. Moreover, it is possibly complicated by the problem of the age asymmetry of the Late Paleozoic glaciations.

INTRODUCTION

The problem of the Northern Hemisphere climate during the Great Gondwana Glaciation is important: first, for a better understanding of the Late Paleozoic climatic and biotic events, and second, for elucidating the scales of and reasons for the asymmetric position of the Permian climatic belts relative to the equator. The asymmetry of the climatic belts is evident (Fig. 1). There are distinct qualitative and quantitative differences in the distribution of the indicators of a cold climate in the Late Paleozoic Southern and Northern hemispheres. In the Late Paleozoic Southern Hemisphere, glacial deposits, both continental and glacial marine, were present throughout the Gondwana mainlands, including the areas located near 30°S in the Permian (*Earth's ...*, 1981; and others). In the Northern Hemisphere, only northeast Asia, which was situated at high latitudes during the Permian, yields evidence of glacial deposits proper: continental (Gusev and Fleishman, 1938; Andrianov and Andrianova, 1962), reworked continental (Khomentovskii, 1948; Kravchinskii, 1969; Pavlov, 1980; Pavlov *et al.*, 1993), and glacial marine sediments (Andrianov and Andrianova, 1962). A wide distribution of season ice marine sediments characterizes this region (Bobin, 1940; Kalugin, 1958; Andrianov, 1966, 1985; Andrianov and Andrianova, 1962; Mikhailov *et al.*, 1970; Ustritskii and Yashvits, 1971; Epstein, 1972a, 1973; Kashik *et al.*, 1990; see also Table 1). The evidence cited to back up these ideas was insufficient in many cases, limited by indications of unsorted tillitelike rocks and erratic fragments, among which striated boulders were rarely found. The characteristics of those boulders, as well as the structural and facies analyses, were not cited, as a rule. Therefore many studies doubt both the ice and glacial origins of the aforementioned deposits (Mezhvilk, 1956; Vikhert, 1957; Bulgakova, 1967; Belozertseva and Silichev, 1973; Umitbaev and Vel'dyaksov, 1977; Ganelin, 1984; Byakov and Vedernikov, 1990, and others). Some authors, pointing to ice-rafted and glacial deposits, seem to have also cast doubt on their genesis, because the number of ice-rafted and glacial units in their successive publications gradually decreased (Table 1).

The Upper Permian Atka Formation in the southern Yana–Kolyma orogenic zone and the adjacent part of the Okhotsk massif are considered key to resolving these ambiguities. This formation is the important marking in this territory and was traced across an area of about 300 × 400 km. The formation is a part of the Gizhiga regional stratigraphic horizon, which correlates to the Kazanian Stage (Ganelin and Kotlyar, 1984; Byakov and Vedernikov, 1990) or the base of the Tatarian Stage of the Permian (Kashik *et al.*, 1990). The lithology of the formation was studied in detail by Epstein (1972a, 1973), who argued for the ice-rafted nature of its genesis. The results of my investigation along the Ayan-Uryakh River and in the upper reaches of the Kolyma River, between Ayan-Uzyakh settlement and the Duskan'ya River mouth (Fig. 2), argue that glacial marine deposits are present in the Atka Formation, in addition to ice-rafted marine sediments.

DIAMICTS OF THE ATKA FORMATION

The formation has a thickness of 400 - 500 m in the studied area and, according to Epstein (1973), is subdivided into three parts: lower (130 - 230 m) and upper (200 - 225 m), consisting principally of diamicts,[1] and middle, composed chiefly of siltstone and mudstone (60 - 130 m). These three parts were named "horizons" by Epstein and "subformations" by others.

The upper diamict horizon of the Atka Formation is rather well-exposed in the riverside outcrops along the left banks of the Ayan-Uryakh and Kolyma rivers and in the area of the Kulu River mouth (Figs. 2 and 3, outcrops 5 and 6). The lower horizon crops out on the same bank of the Kolyma River, 4 km downstream. Diamicts predominate in the lower and upper diamict horizons and constitute 75 - 95% of their sequences. The diamicts, like other rocks of the Atka Formation, are deformed into asymmetric folds and cleavaged. The cleavage intensity increases, evidently, towards the southeast of the studied area.

Diamict Structure

The diamict consists of a dark grey, sometimes almost black, sandy-silty-clay matrix with a small amount of somewhat lighter stones (gravels, pebbles, and boulders) dispersed in the matrix. The patchy appearance of the rock conditions its local names: "patchy shale" or "speckled rock". Among the Atka diamicts, two principal varieties were distinguished: massive and faintly laminated (Fig. 4). Both have a variable matrix composition. According to Epstein, hydromicaceous clayey and siliceous matter is its principal and most constant component, whose total content varies from 30 to 80%. The content of silty and sandy fractions varies even more, from 3 to 30% and from 10 to 40%, respectively. The faintly laminated diamict is usually distinguished by a more clayey matrix, and the massive variety has a more sandy matrix.

The stone concentration in the diamicts, according to visual estimation using the Shvetsov's Tables (1954), usually varies from a tenth of a per cent to 3%. The massive diamicts show an increase in their content, sometimes up to 5 - 10% and very seldom exceeding 15 - 20%. Small pebbles predominate among the stones. Cobbles are rarely encountered; boulders were occasionally found. The largest boulders found in the studied sequence were up to 35 - 40 cm long. The faintly laminated diamict, as a rule, is characterized by smaller content and dimensions of the stones (Fig. 4).

Roundness, Surface, and Composition of the Stones

The roundness of the stones in the diamicts varies noticeably throughout the sequence. As was pointed out previously (Epstein, 1972a), many of the diamicts contain mostly well-rounded clasts (3 - 4 classes on a scale of 5; Khabakov, 1946). From my observations, the stone's roundness may not be as high in other sedimentary members and beds, where semi-rounded pebbles and cobbles (class 2) predominate quantitatively (approximately 70%) (Table 2). The volumetric or weight content of the latter is significantly higher due to the fact that the largest stones belong, as a rule, to semi-rounded phenoclasts (Fig. 5). Well-rounded (class 3) and very well-rounded (class 4) pebbles occur in small amounts (15 - 20% and 0 - 10%, respectively). In addition, variably rounded and then broken pebbles with sharp or dull, newly formed edges are sometimes present (a few per cent). The average roundness of the stones obtained from two units of the massive diamict is 2.2 and 2.3 (Table 2). These figures, as mentioned, seem to be overestimated due to the fact that they are based not on weight or volumetric estimates, but on the frequency of occurrences. Moreover, well-rounded pebbles are easier to extract from the matrix, and thus they are more frequent in the selections.

Stones with subparallel striations of different dimensions on flattened surfaces (Fig. 3) occur at three localities in the massive diamict (Fig. 6). As a rule, the striation is longitudinal, parallel to the long axes of the stones. The length of these features varies from a few millimeters to more than 40 mm, the width is a tenth of a millimeter to 1.5 mm, and the depth is 0.1 to 0.5 mm. In some cases, the striae are not only parallel, and intersect at sharp angles. These striation systems definitely originated before the burial of the clasts in sediment because at places the striae are overlain by remnants of a small-grained matrix filling the scratches and tightly attached to the clast surface (Fig. 6a). The striated surfaces do not coincide with the cleavage planes developed in the matrix, and sometimes they are perpendicular to them. Moreover, the striated surfaces lack the polish and splintering typical of slickensides. There is no break between them and the matrix, and therefore they cannot be considered the result of tectonic displacements. It is evident in some cases that the striation and corresponding surfaces originated under very prolonged abrasion of the surfaces of fixed stones by a variably-grained abrasive. This is indicated by the striated, newly-formed facets that cut some of the previously rounded pebbles (Figs. 6a and 6b). Judging by the form and subparallel orientation of the striae, the abrasive particles moved in a relatively unidirectional motion. The products of the abrasion were swept away and did not accumulate in the striae, as happens during tectonic displacement, creating the typical "splinterness" of slickensides. The above listed features,

[1] In the literature, these rocks are known by their local names "patchy shale" and "speckled rock" (Kalugin, 1958; and others) or by the more common terms "pebbly mudstone" (Epstein, 1972a) and "diamictite" (Byakov and Vedernikov, 1990). Because the term "pebbly mudstone" is not exact enough and the term "diamictite" has various interpretations, I prefer to use the term "diamict" (Harland *et al.*, 1966) to the rock similar to the Atka "speckled rock". This term applies to all clastic sediments and rocks containing large clasts (stounes) dispersed in a relatively small-grained matrix.

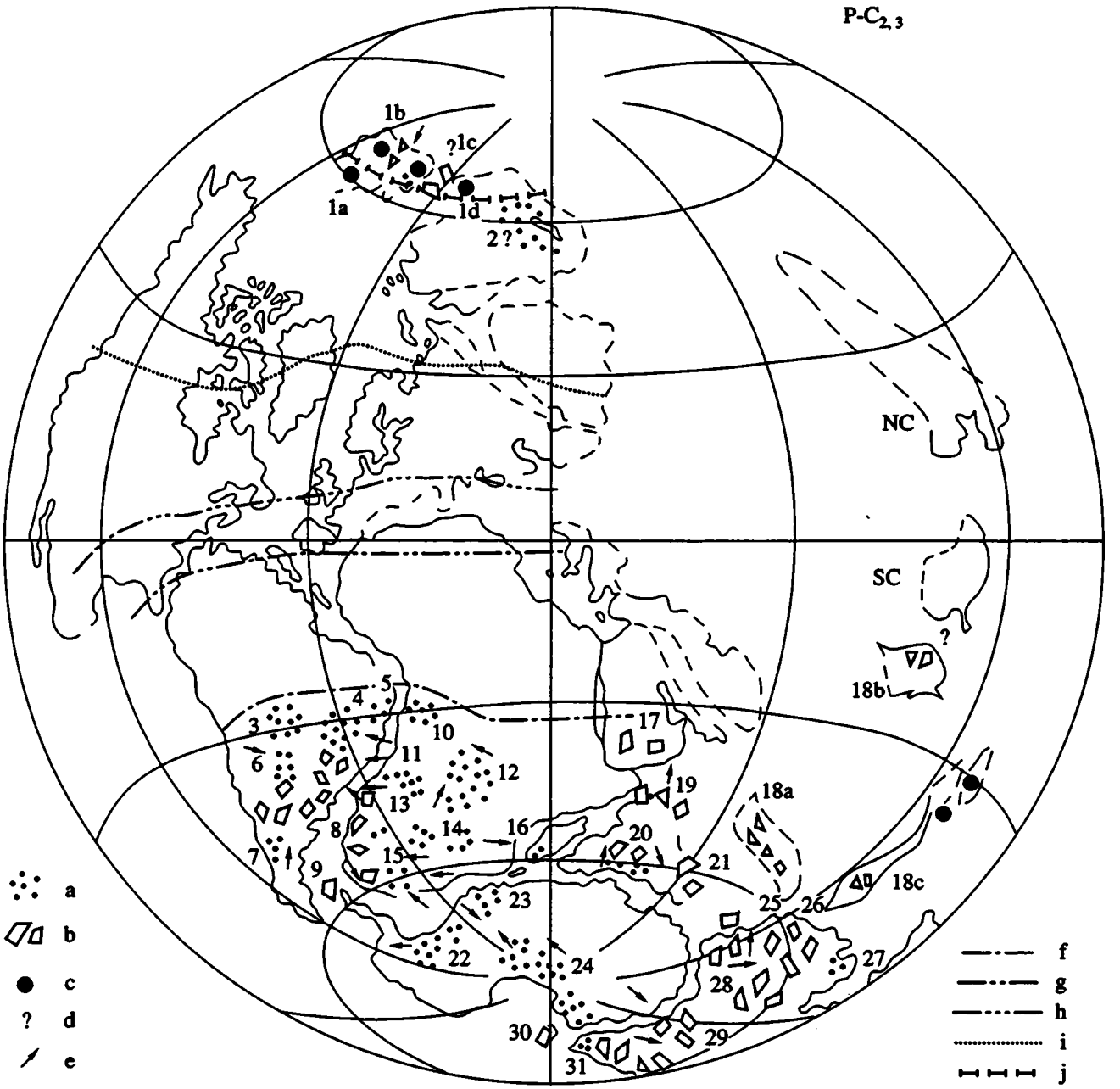

Fig. 1.

especially the newly-formed ("superimposed") facets with systems of subparallel striae, indicate the glacial reworking of the clasts; the abrasion traces of the seasonal ice on rocks and clasts of the littoral zone have a less regular pattern (Dionne, 1973). This can be explained by the shorter (seasonal) duration of each episode of glacial abrasion, the variable direction of the ice motion during each episode, a wave grinding, and the rotation of the abraded objects at the beach during the summer seasons. The stones with superimposed surfaces found in the Atka Formation, before becoming sediment, were probably frozen into the moraine-bearing ice or were pressed into a bottom-moraine and abraded during the prolonged laminar flow of the moraine-bearing ice near the glacier bottom. It is worth noting that the striated pebbles and cobbles were found in the massive diamict, whose stones are characterized by low roundness (the lower and upper parts of unit 7 in outcrop 5 on the northeastern limb of an anticline and the upper part of unit 9 in outcrop 6). This supports the conclusion that the stones with newly-formed abraded surfaces are of a glacial, not ice-rafting, origin.

Clasts dispersed in the diamicts of the Atka Formation, according to Epstein (1972a), have a highly variable composition and were, in general, transported from far away, over a distance of hundreds of

Fig. 1. Asymmetric distribution of the indicators of a cold climate in the Late Paleozoic.

a – continental glacial deposits; b – glacial marine and continental glacial deposits; c – ice-rafted deposits; d – assumed glacial deposits; e – principal directions of glaciers movement; f – northern boundary of the maximum spread of the Upper Carboniferous–Lower Permian Gondwanian glacial deposits (the northern boundary of the southern cold belt); g – boundary of the equatorial humid and southern arid belts at the end of the Carboniferous–beginning of the Permian; h – simultaneous boundary of the equatorial humid and northern arid belts; i – boundary of the northern arid and temperate belts at the end of Carboniferous–beginning of the Permian; j – approximate southern boundary of the northern cold belt. Paleogeographic reconstruction of the continents after Khramov *et al.* (1982) with modifications for eastern and southeastern Asia after Scotese and McKerrow (1990). Supplemented climatic reconstructions after Chumakov (1987). Figures: 1a – assumed ice-marine deposits of Middle Carboniferous and Late Permian age on the Omolon massif and in the middle reaches of the Kolyma River (Ustritskii and Yashvits, 1971; Mikhailov *et al.*, 1970; Kashik *et al.*, 1990); 1b – ice-rafted (Kalugin, 1958; Epstein, 1972a) and glacial marine (this paper) deposits of Late Permian age in the Ayan-Uryakh Anticlinorium and the Okhotsk massif; 1c – Upper Permian ice-rafted deposits and assumed Carboniferous–Lower Permian glacial marine and continental glacial deposits of western Verkhoyansk region (Andrianov and Andrianova, 1962; Andrianov, 1966, 1985); 1d – assumed Middle Carboniferous ice-rafted deposits of the Yudoma region (Bobin, 1940); 2 – assumed Middle Carboniferous glacio-fluvial deposits of the Siberian platform (Pavlov, 1980; Pavlov *et al.*, 1993) and possible reworked products of the Permian continental glacial deposits (Khomentovskii, 1948; Kravchinskii, 1969; Pavlov, 1980; Pavlov *et al.*, 1993); 3 – Jauri Tillite and subdivision 7 of central Brasilia (P-C); 4 – Itarare Subgroup, Parana Basin (C_3-P_1); 5 – Batinga Fm., Sergipe-Alagoas Basin (C_3-P_1); 6 – Tarija (C_2), Charata (C_3), Hoyada Verde (C_2), Ansilta (C_3-P_1), and other formations of the Calingasta-Upsallata, Paganzo, Bajo de Veliz, and other basins; 7 – Tepuel Group (C_2) from the Central Patagonia Basin; 8 – Sause Grande Fm. (C_3-P_1), Argentina and P_2 unnamed beds, Argentine shelf; 9 – Lafonian Tillite (C_2-P_1), Falkland Islands; 10 – N'Khom Group (C_3-P_1), Gaboon; 11 – Lutoe Group (C_3?-P_1), north Angola; 12 – Lukuga Tillites (P_1), Kongo Basin; 13 – Dwyka Group (C_3-P_1) from Namibia and Botswana; 14 – Tilitica Group (C_3-P_1) from Mozambique; Dwyka Group (C_3-P_1) from Zimbabwe; 15 – Dwyka Group (C_2-P_1), Karoo Basin; 16 – Sakoa Group (P), Madagascar; 17 – tillite in the Wajid Sandstone (C_3-P_1), Yemen; Haushi Fm. (C_3-P_1), Oman; 18a – glacial marine Upper Permian–Lower Carboniferous deposits from Tibet (Gao, 1986; Yu Ziye, 1989); 18b – presumably glacial marine deposits of the Early Carboniferous–Late Permian age from the Indosinian massif (Hainan Island: Yu Ziye, 1989); 18c – glacial marine and marine ice-rafted Upper Permian–Lower Carboniferous deposits from the Sinoburman block (*Earth's* ..., 1981; Gao, 1986; L. Frakes, personal communication, 1992); 19 – Tobra Fm. (P_1) of Salt Range; Golabgarh (C_3), Blaini (C-P_1), and other formations from Kashmir and Lesser Himalayas, India; Tinichu and Lachi (C?-P_1) from Nepal and Sikkim; 20 – Talchir Fm. (C_3-P_1), Indostan; 21 – Garu, Ranjit, and other formations (C_3-P_1), Bhutan and Assam; 22 – Whiteout Conglomerate (P_1?), Ellsworth Mts.; 23 – Gale Siltstone (D_1?-P_1), Pensacola Mt.; Beacon Fm. (C_3-P_1), Queen Maud Land; 24 – Buckeye Tillite (P_1), Pagoda Fm. (C-P_1), Darwin Tillite (C-P_1), and Metschel Tillite (P_1) from Transantarctic Mts.; 25 – Lyons Fm. (P_1), Nangetty (P_1), Stockton (P_1) from southwest Australia; 26 – Grant Group (C_3-P_1), Paterson Fm. (P_1) from Canning Trough; 27 – Keep Inlet Beds (P_1) from northwest Australia; 28 – Paterson Fm. (P_1) from Officer Trough; 29 – Boorthanna (P_1), Merrimelia (P_1), Crown Point (P_1), and Purni (P_1) Fms., Cape Jervis Beds (P_1) and others from south Australia; 30 – Wynyard (C_3-P_1?), Stockers (C_3-P_1?), and Zeehan (C_3-P_1) tillites, traces of iceberg rafting (C_3-P_2), Tasmania; 31 – Coleraine Fm. (P_1?), Bacchus Marsh Unit (P_1), Wild Duck Creek Conglomerate (P_1), and others, Victoria; 32 – Joe-Joe (C_3), Cattle Creek (P_1), Ingelara (P_1), and Peawaddy (P_2) fms., Queensland; Rocky Creek Conglomerate (C_2), Spion Creek Conglomerate (C_2), Currabibula ($C_{2,3}$), Kullatine ($C_{2,3}$), Mount Johnstone (C_2), Seaham (C_3), Lochinvar (P_1), Allandale (P_1), Branxtone (P_1), Muree ($P_{1,2}$), Mulbring (P_2), and other formations, east Australia. NC – northern China; SC – southern China.

kilometers from the source area. This author considers the stones' principal source to be the acidic and intermediate Devonian rocks, which are widespread throughout the south, on the Okhotsk massif, and in other regions of the assumed Beringia. Rare clasts of igneous and metamorphic rocks seem to have been transported from the same place. The Kolyma massif, situated to the north, may be the provenance for the clasts of sedimentary rocks, which are present in the diamicts in minor quantities. This allows me to qualify the majority of the stones from the Atka Formation as erratic in a broad sense.

In addition to the clasts of erratic rocks, small angular and tabular clasts of black shale, analogous to those observed in the Atka Formation sequence, sometimes occur in the diamicts. In some cases, small (2 - 5 mm) rounded fragments of other diamicts are found. The latter differ from the host diamict in color, structure, or composition of the included clastic grains. These diamict granules strongly resemble "till pellets", which are typical of recent and some ancient glacial marine deposits (Ovenshine, 1970; Anderson and King, 1981; Young, 1981; Chumakov and Krasil'nikov, 1991, and others), and are formed as a result of the release of small fill lumps when the iceberg thaws. Although it was suggested that similar pellets can be formed in marine ice (Goldstein, 1983), they are more typical of iceberg deposits. Spherical or elliptical carbonate-silty-sandy concretions a few centimeters to a few decimeters in diameter are common in the diamict matrix. Rare remains of marine fauna, brachiopods, gastropods, crinoids, bivalves, foraminifers (Byakov and Vedernikov, 1990), as well as sponge spicules (Epstein, 1972a, 1973) occur in these rocks.

Structures

The massive diamict, forming members from 2 - 3 m to more than 70 m thick, is almost devoid of bedding. Only rare lenses of sandstone, gritstone and small-pebble conglomerate provide a means of determining the bed position. These lenses are, in general, small (a few centimeters thick and 0.5 - 1.5 m long), with a well-defined bottom. They are frequently represented by a gritstone or small-pebble conglomerate at the base and have a gradational contact with the host rock at the top. The faintly laminated diamict sometimes seems to also constitute large lenses in the massive diamict. The thickness of those lenses is up to a few meters, and the length, tens of meters. In addition, isolated oval or elliptical inclusions of sandstone, sometimes layered and spirally or disharmoniously deformed, occur in the massive diamict (Fig. 7c). The dimensions of those

Table 1. Indications of the presence of marine ice-rafted (IR), glacial continental (GC), and glacial marine (GM) deposits in northeastern Asia

Region	Reference	Area	Stratigraphic unit	Indicated age					Assumed genesis			Principal arguments for this genesis
				C_2	C_3	P_1	P_2	Stage	GC	GM	IR	
Verkhoyansk Range	Gusev and Fleishman (1938)	North Kharaulakh	Zh-Horizon			---			+			Polished wavy base
	Bobin (1940)	Southern Verkhoyansk Range, Allakh-Yun' district	Allakh-Yun' Fm.				⊢⊣				+	Diamicts with variably-grained, rounded erratic boulders, fresh feldspars, and cold-water fauna
	Kalugin (1958)	Okhotsk-Kolyma Watershed	Atka Fm.			- -	- -				+	Dispersed boulders and pebbles of various rocks
	Epstein (1972a)	"	"				⊢⊣	Kazanian			+	Dispersed well-rounded erratic fragments in an unbedded silty-clayey matrix; wide distribution
	This work	"	"							+	+	Massive and bedded diamictswith variably-rounded erratic stones, sometimes with glacial striation; dropstones; till pellets
	Andrianov and Andrianova (1962)	Western Verkhoyansk Range	Dulgalakh Fm.				⊢⊣	Upper part of the Kazanian		+?	+	Diamicts, variably-rounded clasts, occasionally striated
	"	"	Echii Fm.			⊢⊣		Artinskian			+	Diamicts with well-rounded erratic fragments
	"	"	Kygyltas Fm.			⊢⊣		Asselian	+	+?	+	Diamicts with variably-rounded erratic fragments; large blocks, varvite, and cold-water fauna
	"	"	Uchagan Fm.	?-	- -				+?		+?	Clay sediments with large blocks
	Andrianov (1966)	"	Dulgalakh Fm., lower Subformation				⊢⊣	Upper part of the Kazanian			+	The same as in Andrianov and Andrianova (1962)
	"	"	Kygyltas Fm.			⊢⊣		Asselian-Sakmarian			+	The same as in Andrianov and Andrianova (1962), with the addition of striated clasts and fresh feldspar fragments
	Andrianov (1985)	"	Dulgalakh Fm., lower Subformation				⊢?⊣	Ufimian? upper part of the Kazanian?			+	The same as in Andrianov and Andrianova (1962)

Table 1 (Contd.)

Region	Reference	Area	Stratigraphic unit	Indicated age					Assumed genesis			Principal arguments for this genesis
				C_2	C_3	P_1	P_2	Stage	GC	GM	IR	
Omolon massif	Mikhailov *et al.* (1970)	Omolon massif					⊢⊣	Kazanian			+	Diamicts with well-rounded clasts
	Ustritskii and Yashvits (1971)	Omolon massif and westermore areas		⊢⊣				Near the Moscovian/Bashkirian boundary			+	Unsorted, unbedded diamicts with variably-rounded erratic fragments; constant characteristics along the layer strike
	Kashik *et al.*(1990)	Omolon massif	Gizhiga Horizon				⊢⊣	Lower part of the Tatarian			+	Diamicts with well-rounded large erratic boulders; paleotemperature; paleolatitude; paleogeography
Siberian platform	Khomentovskii (1948)	Angara–Lena	Base of Tunguska Group						+			Large erratic boulders and pebbles
	Kravchinskii (1969)	Siberian platform and its southeast surroundings	"	- - -	- - -	- - -						Boulder conglomerate; paleofacies; high paleolatitude; paleogeography
	Pavlov (1980), Pavlov *et al.* (1993)	Angara–Lena	Lower Borullou Subformation				- - -		+			Variably-grained sand with pebbles; their wide distribution; glacio-fluvial and pebbles
	"	"	Lapchan Fm.	⊢⊣				Bashkirian	+			Unsorted rocks with exotic pebbles and boulders (moraine?)

Note: Age substantiation: ⊢⊣ based on guide fauna; - - - based on another fauna and stratigraphic position.

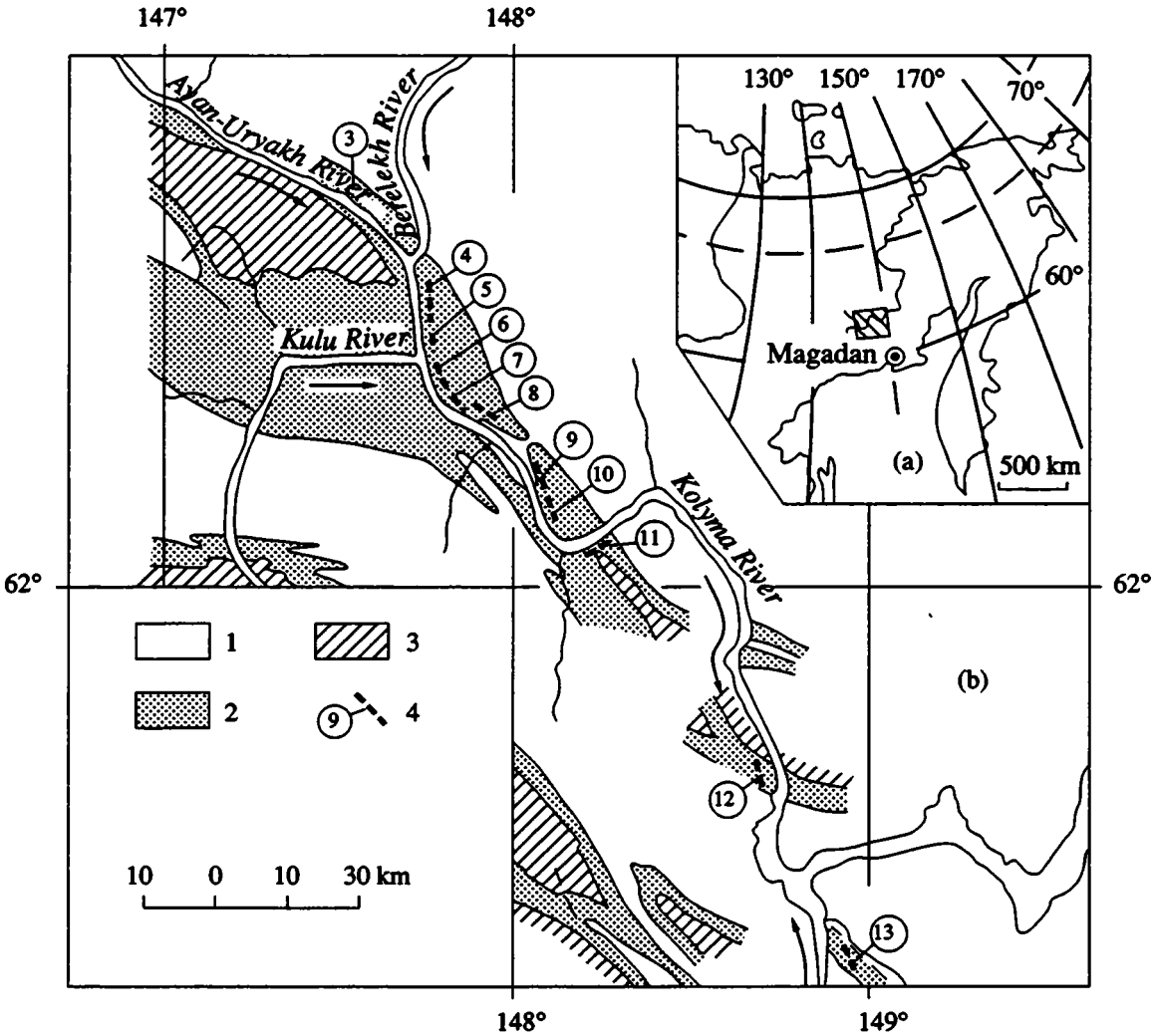

Fig. 2. a – Location map for the studied area; b – distribution of the Atka Formation in the upper reaches of the Kolyma River and the studied outcrops 1 – rocks younger than the Atka Formation; 2 – Atka Formation; 3 – rocks older than the Atka Formation; 4 – studied outcrops and their numbers.

inclusions are 0.5 - 1.5 m (Fig. 7c). Their somewhat uneven surface may be covered by thin coatings of clayey matter and distinctly smoothened. These inclusions very much resemble the olistoliths and slump bodies, which suggests that episodic submarine slumps occurred during the accumulation of the massive diamict. Sometimes, due to the presence of the lenslike interlayers, fragments of disharmonic slump folds can be observed (Fig. 7a). The blocks of shale and faintly laminated diamict found in the massive diamict were also large olistoliths. These blocks are usually slightly deformed or have a layering discordant relative to the general layering of the host massive diamict. The blocks may be few meters thick and a few tens of meters long.

It is more difficult to explain the origin of thin wedgelike or lenslike bodies of gravelly sandstone that cross the units of the massive diamict, often at very steep angles (Fig. 7d). It is conceivable that they are the clastic dikes.

The faintly laminated diamict forms in the layers and units from 3 - 4 to 15 m thick of the upper Atka Horizon, which are distinguished by their darker colors. The alternation of dark clay and slightly lighter silty clay laminae is observed in its matrix. The boundaries between the laminae are obscure, sometimes affected by small-scale bioturbation or traces of loading, and slightly wavy (Fig. 4). The thickness of the clay laminae varies from 1 - 2 mm to 5 mm, and that of the silt laminae, from a tenth of a millimeter to 1 - 2 mm. The laminae have a changing thickness and commonly consist of a series of small lenses. Thin beds (5 - 20 cm) of fine, low-carbonate sandstone occur rarely in the faintly laminated diamict. Fragments of submarine slump folds are also noted (Fig. 7b).

The pebbles dispersed in the faintly laminated, bedded diamict usually deform the underlying laminae,

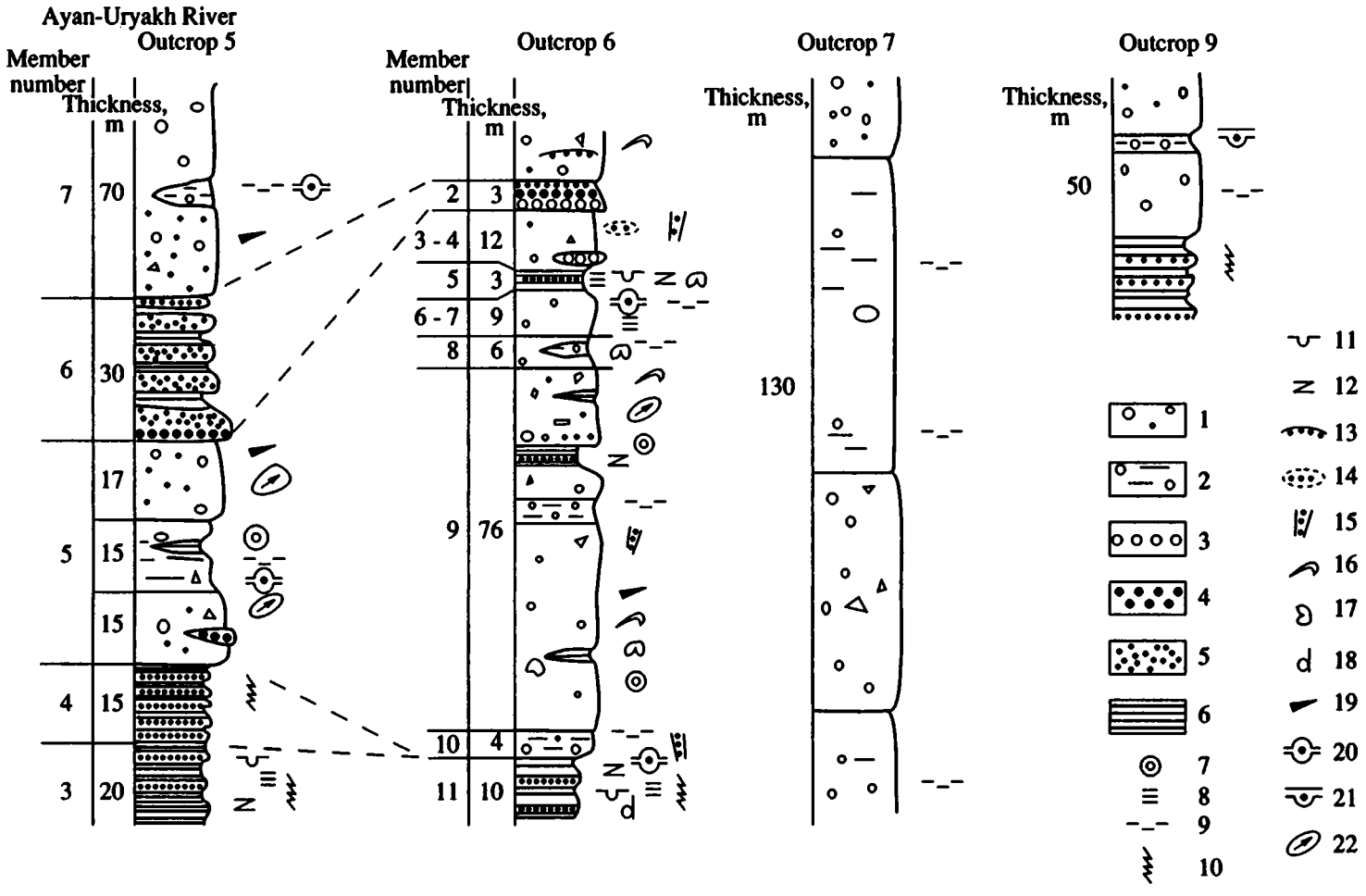

Fig. 3. Characteristic sections of the Atka Formation.

Outcrop 5 (left bank of the Ayan-Uryakh River, 3 km upstream from the mouth of the Kulu River); outcrop 6 (left bank of the Kolyma River, 1.5 km downstream from the mouth of the Bystryi Creek); outcrop 7 (left bank of the Kolyma River, 5 km upstream from the Chainyi Creek); outcrop 9 (left bank of the Kolyma River, 1 km upstream of the Ukhomyt Creek).

1 – massive diamict; 2 – with faint lamination diamict; 3 – conglomerate; 4 – gritstone; 5 – sandstone; 6 – siltstone and mudstone; 7 – carbonate concretions; 8 – thin bedding; 9 – bedding; 10 – rhythmically-bedded members; 11 – load casts; 12 – subaqueous slump structures; 13 – thin lenses of sandstone and gritstone; 14 – large blocks of bedded rocks; 15 – cutting bodies and lenses of grit sandstone; 16 – fragments of sandstone and shale beds with internal plastic deformation; 17 – slump rolls; 18 – water escape structures; 19 – plate fragments of mudstone and siltstone; 20 – lonestones; 21 – dropstones; 22 – boulders and pebbles with glacial striations and secondary facets.

Fig. 4. Faintly laminated diamict from the Atka Formation. White small limestone pebble is seen above (Ayan-Uryakh River, outcrop 4).

whereas the overlying laminae blanket the inclusions in a smoother fashion. In some cases, the details of the interrelations between the clasts and the lamination can be observed. On both sides near the clast bottom, small moundlike heights ("splash structures") are seen; the clast ruptures the underlying laminae and the laminal matter is squeezed upward and outward. The laminae lean against the sides of the clast, while the overlying laminae drape the latter smoothly (Fig. 7e). Such an asymmetry of the deformation above and under the clasts can be explained most easily by their fall onto the surface of fine sediment followed by a subsequent increase in deformation due to loading. The above described clasts can be justifiably considered "dropstones".

ROCKS ASSOCIATED WITH THE DIAMICTS

Among the diamicts of the upper Atka Horizon, members of rhythmically alternating sandstone, siltstone, and, more rarely, shale are present. The thickness of these units varies from a few meters to 30 m, comprising in total from 10 to 20% of sequence of the upper horizon. The thickness of the elementary rhythms in those members varies from 10 cm to 1 m. Their sandstone is characterized by graded bedding and, according to Epstein's definition, has a volcanomictic (1972a, 1972b) or, more precisely, greywacke (1977) composition. The grain size and the thickness of the sandstone layers decrease gradually upsection, and the role of fine-grained rock increases. On the contrary, the sandstone layers in the lower parts of those units commonly have a grit admixture or interlayers of a small-pebble conglomerate. The latter constitute the basal part of the sandstone layers and usually have a clearly identifiable erosional bottom. Thin interlayers with abundant small, angular clasts of the underlying shale are sometimes observed in the lower part of the gravelly sandstone. These interlayers very much resemble "rip-up structures" described by English authors

(for example, Visser, 1983; and others) and are characteristic of turbid, grain, and debris flows. The rhythmical members evidently occur as lenses, because even the largest of them wedge out or decrease greatly in thickness in the neighbouring sequences (Fig. 3, nos. 5 and 6). These members most likely fill in erosional incisions and represent sediments of turbid, grain, and partly debris flows, which accumulate in submarine channels.

Beds and thin members of dark, very fine-laminated (a tenth of a millimeter to a few millimeters) carbonaceous shale and siltstone, rhythmically alternating, are observed at the base of the upper diamict horizon. Thin-bedded, small-grained grey sandstone, forming layers 0.3 - 2 m thick, occur in that shale along intervals of a few decimeters to a few meters. This sandstone is slightly carbonate and is intensively disturbed by traces of loading, small submarine slump folds, and, possibly, water-escape structures. Rare inclusions of granules and pebbles are observed in some shale interlayers. These same shale and sandstones constitute some of the above mentioned blocks in the overlying diamict massive. The shale and sandstone with gravels probably mark the initial formation stage of the upper diamict horizon.

The proportion of shale, sandstone, and gritstone in the studied sections of the Atka Formation increases southeast from the mouth of the Kulu River. In outcrops 9, 10, and 11 (Fig. 3), they gradually take on a more important role and comprise approximately one half of the formation in the latter. Farther to the southeast, on the right bank of the Kolyma River in the area of the former upper Kolyma Rapids (outcrop 12, Fig. 3), a schistose diamict with rare, small, and poorly rounded pebbles seems to predominate once again.

ORIGIN OF THE ATKA DIAMICTS

The remains of marine fauna, predominantly fine and sometimes thin-bedded matrix, traces of subaqueous slumps, and co-occurrences with graded-bedded rocks and evidently with erosional channels, are all evidence indicating the marine and deep-water origin of the Atka diamicts. At the very least, one can say confidently that they accumulated in calm waters, where the clay was able to settle almost constantly, the seafloor slopes were steep enough for the formation of the gravity flows, and the relief necessary for the incision of these flows into sediment existed. Small lenses of gravelly sandstone grading into the host diamict indicate weak bottom currents that washed fine particles out of the sediment or suspension. The wide destribution of relatively regularly dispersed erratic stones (up to small boulders) and typical dropstones in the diamicts is evidence that they were transported a considerable distance from the shore areas of the basin by some powerful and regularly acting geological agent that, at the same time, did not disturb the calm hydrodynamic conditions necessary for the deposition of the fine sediment that constitute a significant part of the diamict matrix. The submarine slumps could not have been those factors, as was proposed by some studies on the diamict of the Gizhiga Formation from the Omolon massif (Ganelin, 1984 and others). The slumps would have disturbed the layering; however, this is only occasionally observed in the Atka diamict sequence in some units. Other gravity flows, including turbidity flows, which, according to Belozertseva and Silichev (1978), deposited the "speckled rock" of the Menkechen Formation in the south Verkhoyansk Range region, could also not have been the agents under question. The gravity flows do not disperse stones in a sedimentary section but, rather, create a graded bedding with large clasts concentrated in separate layers. Moreover, the gravity flows, with the exception of slumps, do not form thick layers (Visser, 1983) analogous to those noted in the Atka diamicts, and, as a rule, they erode the underlying fine sediments. Erosion was not observed at the base of the diamict units in the studied sequences. The turbidity currents are not capable of carrying fragments larger than small pebbles (Heezen and Hollister, 1963). We mentioned above that traces of gravity flow activity, through present in the Atka Formation, are restricted by discrete units mainly comprised of layered rock and, more rarely, diamict.

It is unlikely that the occurrence of the stones in the fine sediments resulted from volcanic eruptions, because typical volcanic bombs and lapilli were not found among the stones, and even proponents of this theory only rarely noted volcanic ash in the diamict matrix (Byakov and Vedernikov, 1990). Furthermore, not all the stones have a volcanic origin, and some of them are rounded. Epstein (1972a, 1972b) correctly noted that this contradicts the theory claiming their explosive origin. A direct correlation between the grain size and thickness of the layers in the Atka Formation is also uncharacteristic of the effusive-sedimentary deposit (Epstein, 1977).

The assumption that the stones could have been transported to the Atka basin by driftwood, algae, or marine animals is unlikely for several reasons. First of all, biogenic transportation over a long distance occurs on a rather limited scale. The most powerful agent of biogenic transportation, driftwood, seldom carries stones and usually loosens most of them when the soil captured by roots is soaked. This is why the dispersed pebbles and boulders are rare even in coal seams, although abundant trees brought by water contributed to their formation. Usually, a few or several dozen stones are found per coal field or several hundreds stones per group of coal-fields. Therefore, special publications (Price, 1932; Stutzer, 1940; Turlei, 1941; Dix, 1944; Karlov, 1956, 1963; and many others) are often devoted to those findings. Detailed investigations of some of the most well known localities containing boulders dispersed in coal seams leads us to the conclusion that their transportation was accomplished in two stages. First, transportation by rivers over a long distance (Karlov, 1956) and possible multiple redepositions in different environments, probably including the

littoral zone.[2] The second stage was marked by the local transportation of the stones in the roots of trees within the coal basin.

Rare cases of stone transportation over great distances and even fine earth are known from navigators' observations (Emery, 1955), but the scale of this event is negligible. For example, on the coral islands of the Pacific, the solid, erratic stones transported in such a manner were highly valued and were considered the property of the chiefs (Darwin, 1860), which indicates the great rarity of those findings. The quantity of stones brought by driftwood over an observation period of 3 years on 10 Californian beaches was estimated to be about 4300 (Emery, 1955). This means that even if overestimated, they can not exceed 0.01% of the beach sediments.[3] Taking into account that a significant part of the driftwood with stones is washed out again from the beaches to the sea during storms, the real figure would be several times less.

The volume of the clastic material carried by sea algae is also small. The quantity of clasts carried by them under rather favorable hydrodynamic conditions is estimated, based on the dredging data, to be significantly more than 600 per sea floor square kilometer (Shumway, 1953). Even if we were to assume their real concentration to be two orders more than indicated, they would constitute only about 0.005% of the sediments. The scale of the stone transportation by modern animals (mostly of gastroliths by Pinnipedia) is thought to be either insignificant and local (Emery, 1941) or regionally limited and typical of relatively small areas of islands and coasts (breeding-grounds) (Fleming, 1951). A longer transportation distance of the gastroliths was supposed for some of the Cretaceous dinosaurs in Western Europe (Hawkes, 1951 and others), however, even in this case, the number of stones is measured on a scale of hundreds per region.

After comparison with the above cited figures, the total volume of the stones dispersed in the Atka diamicts, even according to the minimum estimates,[4] seems to be huge, about 250 - 500 km^3. The sandy fraction, which, according to Epstein (1972a), chiefly came with the stones, can possibly be added to that. In my opinion, this suggestion is supported by the presence of dispersed sandy grains, including those of very coarse size, in the silty-clayey matrix of the faintly laminated diamict. Taking the sandy fraction into consideration, the total amount of the clastic material in the diamicts is not less than 3600 km^2. However, it should be taken into account that part of the silty-clay material from the diamicts also came to the Atka basin together with stones and sand.

The duration of the second half of the Kazanian or the first half of the Tatarian can be estimated to have been about 2.5 Ma (Harland *et al.*, 1990). On the basis of the above figures, one can deduce that the input of the erratic terrigenous material to the Atka basin was not less than 1400000 m^3/year. This figure is comparable, as a first approximation, with the amount of the terrigenous material that, according to recent data, is rafted by the seasonal ice in the Sea of Okhotsk (1756000 m^3/year; Stepanova, 1985), on the Murmansk coast (4500000 m^3/year; Tarasov, 1981), and in the Beaufort Sea or the Fox Basin (1000000 m^3/year and 2000000 - 4000000 m^3/year, respectively; Drewry, 1986). However, a more detailed consideration shows that the supply of erratic material to the Atka basin was even higher than ice rafting, for example, in the Sea of Okhotsk or the Bering Sea (Lisitsyn, 1961, 1966). The recent estimations cited above apply to all the terrigenous material captured by ice in the littoral zone; the estimations were made based on the content of the material inside the ice and on its surface. Clay and silt constitute a significant part of the material. However, it is impossible to define what portion of the Atka Formation consisted of silty-clay material transported with the erratic matter. On the other hand, only a part of the ice-captured material is carried away outside the coastal zone, whereas the other part is redeposited inside this zone (Drewry, 1986). Thus, the figure obtained for the Atka Formation is evidently underestimated, because it corresponds to only a part of the total sedimentary load of the fast ice. Finally, the scale of the ice rafting must be proportional to the length of the basin's shoreline, and the maximum assumed length of the Atka basin shoreline was several times shorter than that of the Sea of Okhotsk. Thus, it can be concluded that the intensity of the transportation of the erratic material to the Atka basin was several orders of magnitude higher than biogenic transportation and seems to have been significantly more intensive than typical seasonal marine ice rafting.

The second objection against the idea that biogenic transportation played an important role in the Atka basin is that, normally, a greater part of the floating wood, algae, and animals' corpses drift, pushed by the wind, to the nearest shores and is cast onto the littoral; the clasts captured by them are buried in the shallow-water facies of a dynamically active zone. As for the Atka diamicts, it has already been noted that they accumulated in calm waters and, most probably, at a distance from the coast. Therefore, it is difficult to account for the large volume of erratic material in the Atka Formation using biogenic transportation.

Lastly, only the clasts of a certain size, usually from large pebbles to small boulders, can be captured and transported by marine organisms (Dalland, 1976), while erratic granules and coarse sand play an important role

[2] A high level of roundness and quartz or quartzite composition of many of the boulders indicates this. The stones found in roots of driftwood on the modern beaches have an average roundness of only about 0.9 according to Khabakov's scale (Emery, 1955).

[3] Proceeding from the following data: the average observed diameter of the stones is 4 cm; assuming (evidently underestimated) the dimensions of each beach to be 10 m × 100 m = 1000 m^2; and the studied beach deposits to be 5 cm thick).

[4] The distribution area of the Atka Formation is approximately assumed to be a triangle 280 × 400 km (about 56000 km^2); the average thickness of the diamicts is about 300 m; the average stone concentration is 1.5 to 3%; and sandy fraction constitutes, on average, about 20 - 30%.

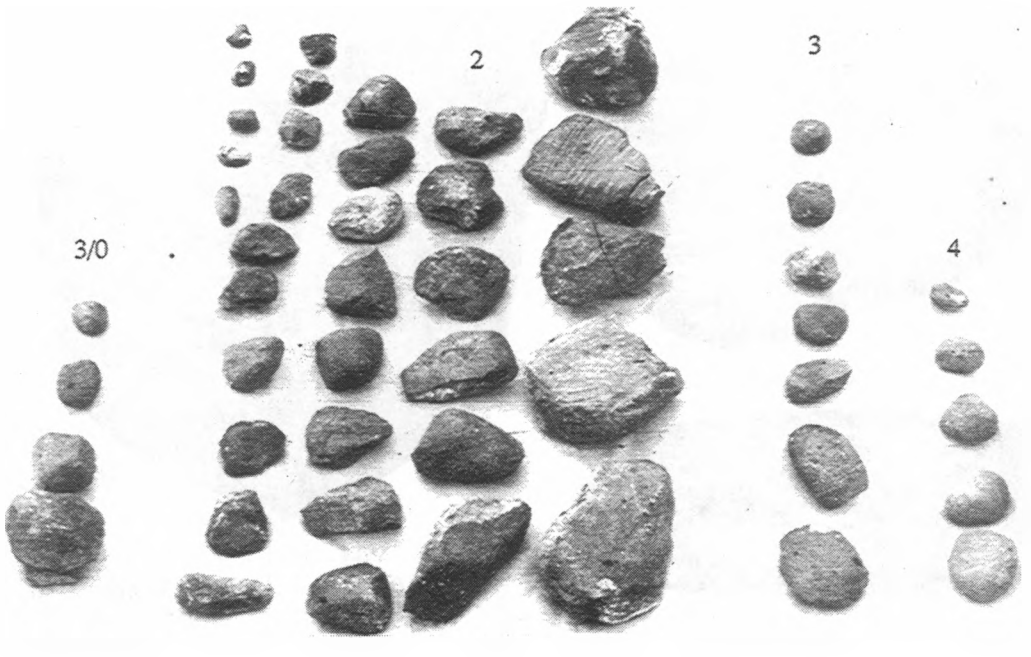

Fig. 5. Distribution of pebbles and boulders from diamicts according to their roundness (figures above are the roundness classes after Khabakov, 1946); Kolyma River, the upper Atka Subformation, outcrop 6, member 9. The largest stone is 11 cm long.

in the Atka diamicts. In addition, the remains of the epifauna characteristic of the biotopes of the algal thickets and its traces in the form of borings are observed in many stones carried by algae (Shumway, 1953). Epifauna and stone borers were widespread in the Permian (Gekker, 1959), also including the Verkhoyansk basin (Lungersgauzen, 1970), however no traces of life activity were observed on the stones from the Atka diamicts.

Comparing the various means of transport of the erratic material over long distances allows me to agree with those researchers who concluded that the coarse clasts were brought to the Atka basin by floating ice. The only question worth pursuing is the nature of the ice that transported the stones. The idea that the diamict units with well-rounded stones were formed under the influence of the seasonal marine ice seems to be valid. At the same time, the low-grade roundness of the stones in other units (averaging 2.2 - 2.3, Table 2) and, above all, the presence of pebbles with clear traces of prolonged glacial abrasion (Fig. 6) indicates that a part of the Atka diamicts was formed by icebergs, in addition to the seasonal ice. The presence in the diamicts of the inclusions resembling till pellets supports this supposition. This part of the diamicts can be termed iceberg (distal glacial marine) deposits. The presence of icebergs in the Atka basin suggests the existence of gla-

Table 2. Roundness of stones in the massive diamict containing striated pebbles and boulders

Sample location	Total number of stones	Distribution of stones according to their roundness (in classes according Khabakov's scale, 1946);							Average roundness in selection
		0	1	2	3	4	Broken stones		
							3/0* (Footnote)	3/1* (Footnote)	
Outcrop no. 5 (member 5), Ayan-Uryakh River	28	–	–	20	6	–	–	2	2.2
	100%	–	–	72	21	–	–	7	55.4
Outcrop no. 6 (member 9) Kolyma River	50	–	–	34	7	5	4	–	2.3
	100%	–	–	68	14	10	8	–	57.5

* The roundness of the broken pebbles is indicated by a fraction, where the numerator is their initial roundness and the denominator is the roundness of newly-formed edges. The general roundness of the broken pebbles is adopted as the mean value between the numerator and the denominator.

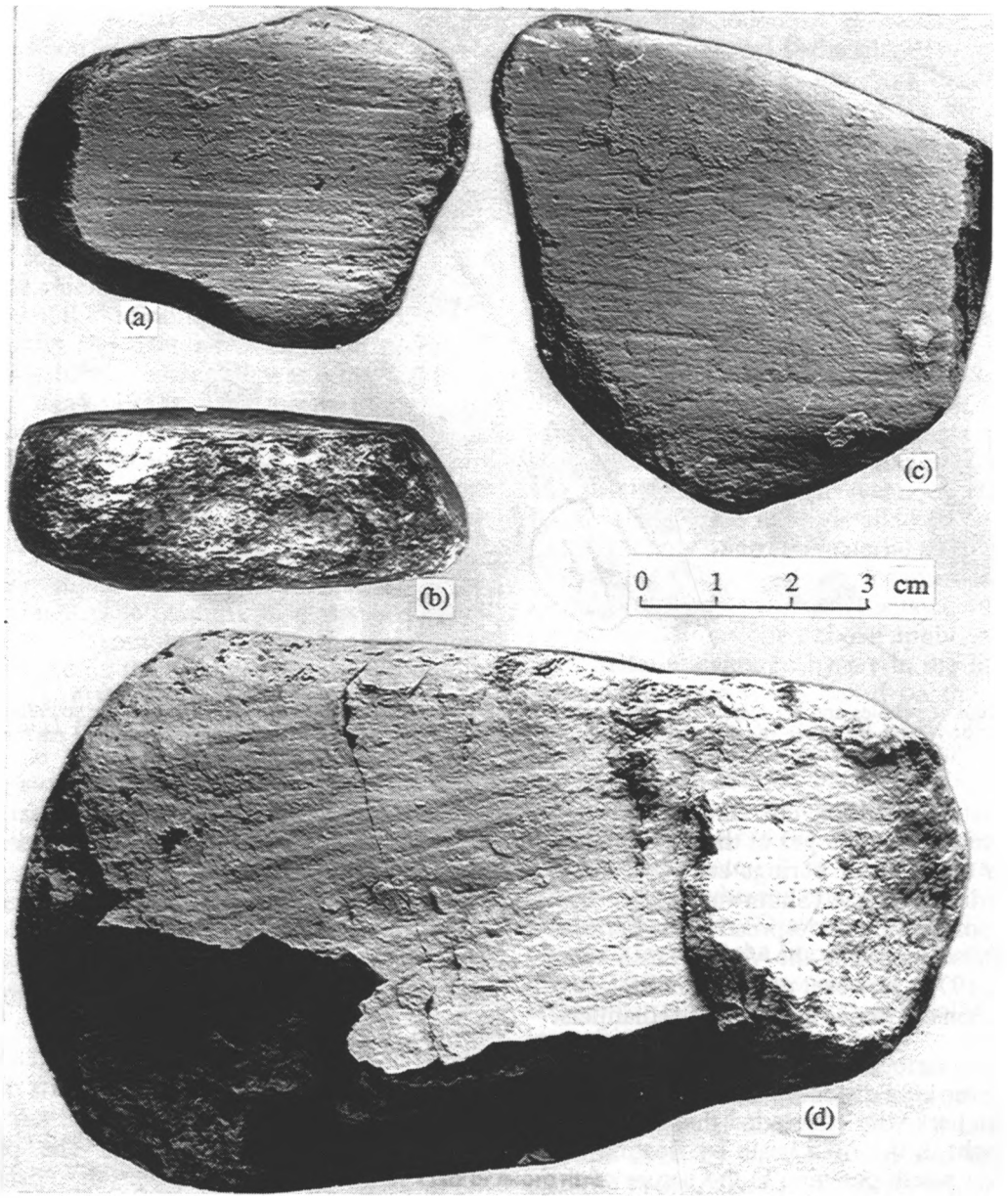

Fig. 6. Glacial pebbles and cobble from diamict, upper Atka Horizon.

a, b, c – Ayan-Uryakh River (outcrop 5, member 5): a – a facet on pebble of aphanitic volcanic rock with a longitudinal and sublongitudinal system of variably-sized striae; in the center and below there are small (few millimeters across) remains of a matrix tightly attached to the surface of the facet. Some striae overlain by matrix are visible; the same is seen in Fig. c. b – the same pebble as in Fig. a, side view; above, a striated facet; note that this facet is superimposed, truncating the semirounded original surface of the pebble. d – Kolyma River (outcrop 6, member 9); small boulder of porphyrite with longitudinal subparallel striation going below the matrix remains (on the right). Photo by A.I. Nazarov and A.A. Okunev.

ciers flowing down to the sea. They apparently were the outlet glaciers, as one must assume that the fast ice regularly formed along a significant part of the coast, and that this ice collected the clastic beach material. Locally, closer to the coast, the shelf glaciers have obviously existed. The lithological peculiarities of the sediments under consideration corresponds to the supposition that the subpolar outlet glaciers with a "warm" bottom participated in their deposition. The bottom streams of melted ice typical of "warm" outlet glaciers could have supplied a large quantity of suspended minerals to the basin, and icebergs may have brought unsorted morainic material, including coarse clasts. The mixing of seasonal ice and glacial materials may be the explanation for the higher average roundness (2.2 - 2.3) of the pebbles in the Atka iceberg diamict in comparison to the glacier sediments, where the average roundness varies between 0.22 and 0.76 and reaches 0.9 only rarely (Epstein, 1986). At the same time, the average roundness of the Atka iceberg diamict

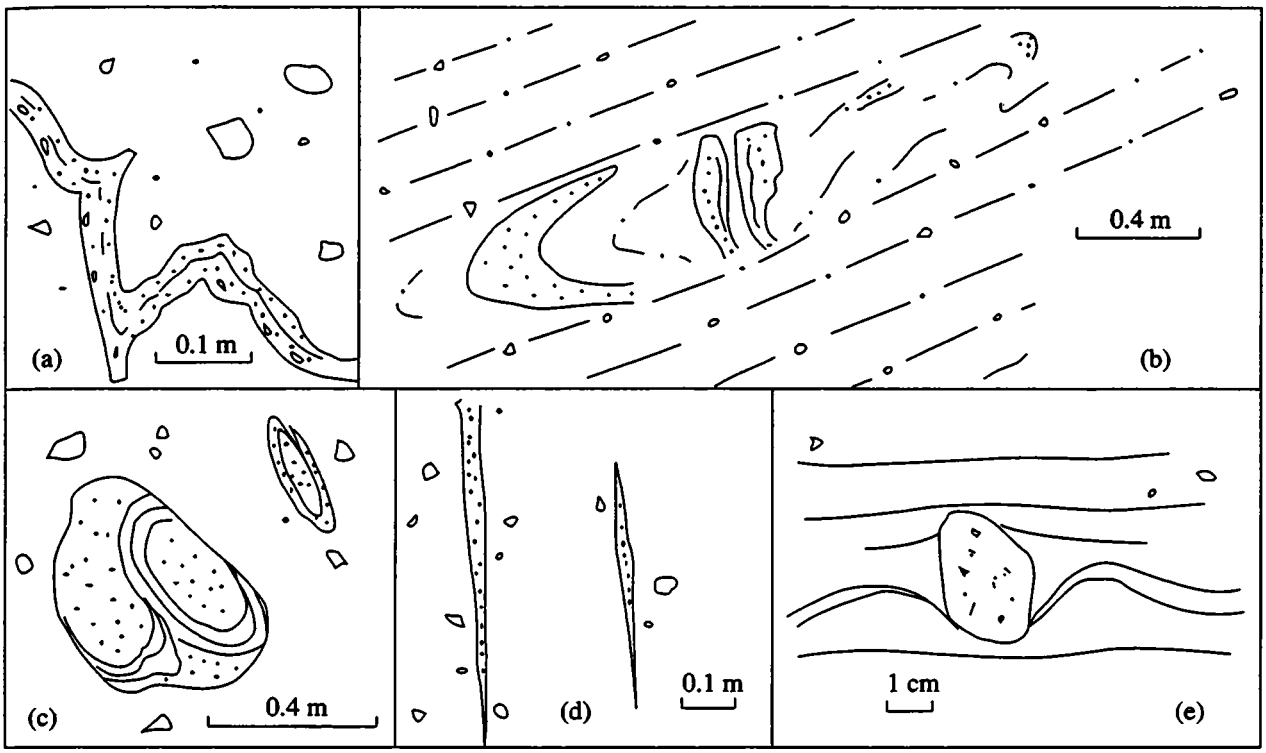

Fig. 7. Typical structures in diamicts of the upper Atka Subformation.
a – Subaqueous slump fold in massive diamict (Ayan-Uryakh River, outcrop 5, member 7); b – sandstone bed with subaqueous slump structures in faintly bedded diamict (Ayan-Uryakh River, outcrop 4); c – slump sand rollers (small olistolites in massive diamict (Kolyma River, outcrop 6, member 9); d – cutting wedge-like and lenslike bodies of grit sandstone in massive diamict (the same locality); e – dropstone in bedded diamict from the Atka Formation; "splash structure", the squeezed underlying beds; beds leaning against stone sides and draping overlying beds are seen (Kolyma River, outcrop 9). See Fig. 3 for legend.

is appreciably lower than that of the ice-rafted diamict from this formation. The average roundness of the latter, as was already noted, reaches 3 - 4 (Epstein, 1972a). The influence of the "warm" outlet glaciers also explains a higher terrigenous material supply of the Atka basin, because such glaciers are capable of transporting a great volume of clastic sediments (Drewry, 1986). The high rate and spatial irregularity of this material' result, as is known from the study of young and ancient glacial marine deposits (Eyles *et al.*, 1985; Powell, 1990; Chumakov, 1992, 1993), in the formation of submarine fans near the head of the glacier with erosional channels and canyons, in frequent slumps of sediments, downslopes of accumulative and erosional seabottom relief, and in the origin of gravity flows usually channelized by those erosional incisions. The Atka diamicts, judging by their fine matrix, were formed at the periphery of such submarine fans. The lenses of the graded-bedded sandstone and silty pelite in the diamicts most likely represent debris flow sediments that filled the channels along the periphery of fans. Some unlayered massive diamicts were probably partly formed by shelf glaciers that floated and even rested on the seafloor (submerged), glaciers, whose emplacement could be to some extent responsible for slump and glacial(?) deformations. Weak bottom currents observed at some places of the modern glacier discharge suspend silty-pelitic particles or, more seldom, wash them out from sediments (Eyles *et al.*, 1985). This results in sediments enriched in coarse clastics ("residual paratills"; Anderson *et al.*, 1982). The above described gravelly-sandy lenses in the massive diamict are most likely of such a nature.

THE PALEOCLIMATIC SIGNIFICANCE OF THE ATKA DIAMICTS

As was noted at the beginning of this paper, the occurrence of glacial sediments in the Upper Paleozoic deposits of northeastern Asia has already been suggested in the literature (Khomentovskii, 1948; Andrianov and Andrianova, 1962; Pavlov, 1980; Pavlov *et al.*, 1993). Some others have not ruled out that possibility (Epstein, 1973). However, glacial marine deposits of the Atka Formation are the first finding of glacial deposits in the Late Paleozoic Northern Hemisphere that has been confirmed by reliable diagnostic indices. The finding suggests that in the middle Late Permian in northeastern Asia, land glaciers existed and reached sea level at some places, giving rise to icebergs and possibly to shelf glaciers. The petrological composition of the stones, southward replacement of the clay diamict by the silty diamict, and a simultaneous increase in sandstone and gravelstone content (Epstein, 1972a,

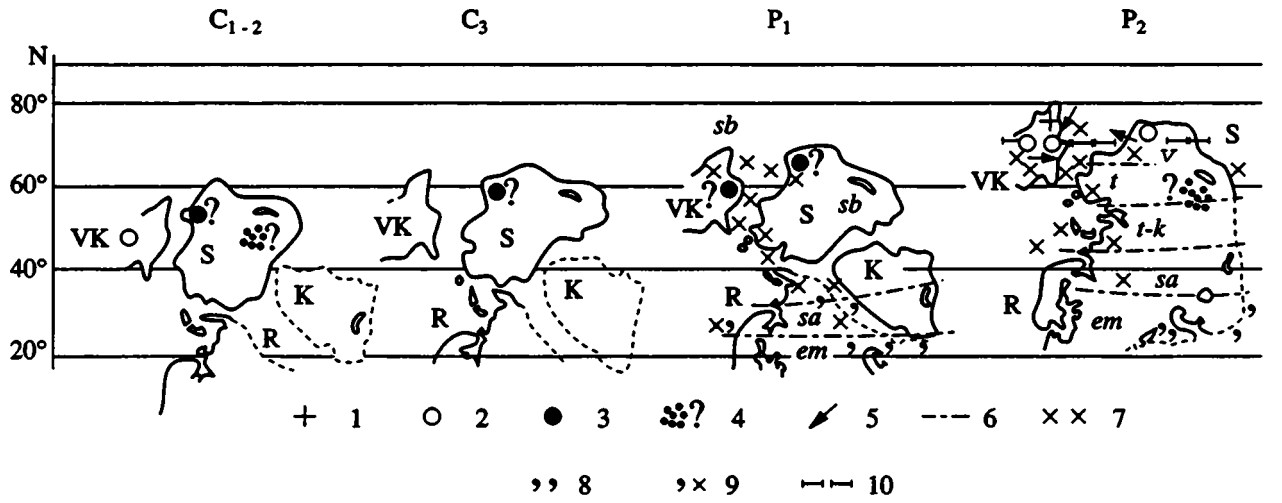

Fig. 8. Scheme of the longitudinal drift of the continental blocks of Northern Eurasia during the Late Paleozoic and some elements of palaeobiogeography (After Khramov *et al.*, 1982; Durante, in press; Ustritskii, 1967, 1993; Waterhouse and Bonham-Carter, 1975).

1 – Seasonal ice and glacial marine deposits of the Atka Formation; 2 – other indications of seasonal ice marine deposits; 3 – indications of continental glacial and glacial marine deposits; 4 – glacio-fluvial (?) and reworked glacial (?) deposits; 5 – assumed direction of transportation of glacial material; 6 – boundaries of phytochorias; 7 – locations of boreal marine fauna; 8 – locations of the Tethian marine fauna; 9 – locations of mixed fauna; 10 – approximate boundary of the cold zone; VK – Verkhoyansk Range–Kolyma continental block; S – Siberian platform; R – Russian platform and the Urals; K – Kazakhstan and West Siberia. Phytochorias: *em* – Euramerican region; Angara Kingdom: *sa* – Subangara region; *sb* – Siberian Region; *t-k* – Taimyr–Kuznetsk subregion of Siberian region; Tunguska–Verkhoyansk subregion of Siberian region: *t* – Tunguska district; *v* – Verkhoyansk district.

1973) indicate that the Okhotsk massif was the principal center of the Atka glaciation. Along with other data, these facies and well-known biogeographical relations (Fig. 8) between the Verkhoyansk zone and Okhotsk massif prevent us from considering the latter as a terrain that drifted to northeastern Asia in the Mesozoic (Zonenshain and Kuz'min, 1993). The Kolyma massif was possibly a secondary center of ice rafting during the Atka time (Epstein, 1972, 1973).

The above data shows that the climate in northeastern Asia at the end of the Permian was colder than it was earlier supposed. The climate, at least sometimes, was glacial rather than cold-temperate, as suggested previously when the Atka diamicts were thought to be completely of an ice-rafting origin. This places the southern boundary of the Northern Hemisphere cold climatic belt for that time in the southern parts of the Aldan shield and the Verkhoyansk–Kolyma zone (Figs. 1 and 8). Such a position is in accordance with most global and regional paleomagnetic reconstructions and paleobiogeographical data. These reconstructions show northeastern Asia and the Siberian platform situated in high latitudes, between 50° and 80°N during the Late Permian (Khramov *et al.*, 1982; Green *et al.*, 1986; Scotese and McKerrow, 1990; Ustritskii, 1993, and many others); they were the northernmost landmasses on the Earth. According to Khramov's reconstruction, they faced toward the North Pole with their modern southern and eastern edges, respectively, and therefore the areas of the upper reaches of the Kolyma River, the Okhotsk massif, and the southeastern part of the Aldan Shield were nearest to the Pole (75 - 80°N; Fig. 8). In this respect, the occurrence of a thick unit of seasonal ice and glacial marine deposits during the Permian seems natural.

As for paleobiogeographical data, as early as the 1930s, many researches noticed a boreal appearance of the Permian marine, especially benthic fauna in northeastern Asia, as well as its specific similarity to the Gondwanian faunas of the glacial regions (Ustritskii, 1967, 1974, 1993; Waterhouse and Bonham-Carter, 1975; Leven, in press, and others). Some elements of the tropical (Tethian) fauna are present at the lower paleolatitudes of Eurasia. The Tethian fauna totally predominates south of the paleolatitude of 20 - 25°N (Fig. 8).[5] The Permian climatic zonality in northern Eurasia is

[5] V.G. Ganelin (1990) correctly pointed out that the replacement of the Tethian fauna with the boreal one during the Carboniferous in northeastern Asia coincides with the change in sedimentation pattern (the shallow-marine carbonate facies is replaced by terrigenous deep-water facies), and that these changes were not simultaneous in different regions. Therefore, he suggests that the distribution of the Tethian and boreal faunas in the Late Paleozoic was conditioned by the depth of basin rather than by latitudinal climatic zonality. This explanation seems valid because a sufficient increase in the depth of basin always results in a decrease in the near-bottom water temperature and faunal changes. Both of these events must have been especially pronounced during the glacial periods, when a thick layer of the cold, near-bottom water, the psychrosphere, arose in the ocean. Several glacial periods are noted for the Late Paleozoic. Tectonic subsidence at different times in different regions, the difference in the paleolatitudinal locations of the latter, and repeated climatic fluctuations could have easily resulted in asynchronous faunal changes. However, this is not sufficient grounds for denying the general dependence of the boreal and tropical benthic faunas distributions on climatic factors.

still clearer from the phytogeographical data. There, in the Late Permian, north of the tropical Euramerican paleofloral region, four successive phytochorias of different ranks accompanied by the corresponding impoverishment of the floras are recorded (Durante, in press; see also Fig. 8). At the same time, it was convincingly shown that although the climate of the Siberian paleofloral zone, combining the three northernmost phytochorias, was temperate, the northermost Verkhoyansk paleofloral region was characterized by a colder climate; thermophilic floras of the Subangara paleofloral region did not penetrate there even during warmings. The latitudinal arrangement of a number of the phytochorias within a single climatic belt alone indicates a strong meridional floral differentiation and, possibly, a high meridional temperature gradient in the Late Permian.

Thus, the sedimentologic, paleomagnetic, and zoo- and phytobiogeographic data, considered seperately and in ensemble, show that the studied area belonged, at least sometimes, to a cold climatic zone in the mid-Late Permian.[6] Discrete, thin diamict layers similar to ice-rafted sediments occur in the Upper Permian sequence in the Kolyma-Okhotsk region both below and above the Atka Formation (Epstein, 1972, 1973). They evidently indicate several periods of climate deterioration during the Late Permian. In addition, more prolonged climatic deteriorations are recorded for the Siberian paleofloral region at the beginning of the Middle Carboniferous and at the beginning of the Early Permian (Durante, in press). The first of them seems to have been especially strong, and the so-called post-lepidophytic flora with a "tundra" appearance was widespread at that time. It is noteworthy that the three climate deteriorations recorded for northeastern Asia in the Permian–Carboniferous correspond, approximately, in age to the last three glaciations of the Gondwanian Glacial Era, which are clearly represented in the Southern Hemisphere (Chumakov, 1985, 1987; Veevers and Powell, 1987). The coincidence of climatic minima in opposing hemispheres is, in itself, noteworthy, and deserves more attention. Furthermore, this raises questions concerning other indications of the presence of ice-rafted and glacial deposits in northeastern Asia (especially in the Middle Carboniferous, uppermost Carboniferous–Lower Permian, and middle Upper Permian).

The above listed data on the Upper Permian glacial deposits in the Kolyma River region suggest that the climatic asymmetry, relative to the equator in the Upper Paleozoic, was somewhat less than was earlier suggested. At the same time, it should be emphasized that the occurrence of glacial sediments in the Atka Formation does not solve questions concerning the spatial climatic asymmetry in the Upper Paleozoic time, because the scale of the climatic deteriorations in the Northern Hemisphere is thought to be significantly less than that in the Southern Hemisphere. Even if all existing indications of Upper Paleozoic glacial and ice-rafted deposits in northeastern Asia were confirmed, one could only address the Upper Paleozoic ice caps and ice sheets of the polar type because, according to paleomagnetic data, the region under consideration was situated at rather high latitudes, about 60 - 80°N (Fig. 8), during the Permian. It should be emphasized that in the Southern Hemisphere, glaciers sometimes reached 30°S during that period.

The principal reason for Late Paleozoic climatic asymmetry is most likely explained by the asymmetrical arrangement of the continents relative to the equator at that time. During the Late Paleozoic, at the middle and partly high latitudes of the Northern Hemisphere, parts of northern Asia and the northern margins of America and Europe, still separated and surrounded seas, were situated for a long time. In the Southern Hemisphere, the immense Gondwana was a continuous landmass from the equator to the pole (Fig. 1). The parallelism between the asymmetries of the glaciations and the continents is also observed in the Late Ordovician and Pleistocene (Chumakov, 1984). In the first case, this was enhanced by the absence of land at the high northern latitudes, and in the second case, was partly masked by the ocean surrounding Antarctica that restricted the expansion of glacial sheets. One can suggest that the albedo of the oceans and seas, which is lower than that of land, the restrictions created by those oceans and seas facilitating the spread of glacial sheets, and favorable conditions for intensive circulation and heat exchange through the hydrosphere, which contained warm belts, were the principal reasons for why the hemisphere, with a relatively large area covered by the oceans and seas in the high and middle latitudes, had a warmer climate.

If the opinion that the Atka Formation belongs to the Tatarian was proven valid, then the issue of spatial climatic asymmetry would be complicated by the problem of the stratigraphic asymmetry of the glaciations, due to the fact that the last episodes of the Gondwana glaciation in the Southern Hemisphere are usually dated by the Kazanian. Although some of the researches do not exclude the possibility that the last glaciers in southeastern Australia existed until the Tatarian (Herbert, 1981; Packham, 1981), others seriously doubt this (Dickins, 1985). It is possible, of course, that the traces of the Late Permian glaciation in northeastern Asia are related to the drift of the continental blocks 10° - 12° northward (Khramov *et al.*, 1982; Scotese and McKerrow, 1990) rather than to the phase of general climate deterioration on the Earth. In addition to the shift of land toward a colder zone, the glaciation development may have been enhanced by orogenesis on the northern active plate margins.

[6] Unfortunately, because of large discrepancy between the data of different scientists and the data on different faunal groups, it is difficult to use the paleotemperature estimations for the Permian basin in northeastern Asia obtained by the Ca-Mg methods (cf. Astaf'eva-Urbaitis and Yasamanov, 1986 and Kashik *et al.*, 1990).

CONCLUSION

The conclusions concerning the origin of diamicts of the Atka Formation cannot be universally applied to all of the locations of assumed Upper Paleozoic glacial and ice-rafted deposits in northeastern Asia (Fig. 1; Table 1). The question concerning the origin of these deposits must be resolved region by region on the basis of a complex set of genetic indications. The works devoted to the Lower Carboniferous Krestyakh conglomerate ("Zh"– Horizon of Gusev and Fleishman, 1938) that have categorically proven its non-glacial origin (Bulgakova, 1967 and others) are a good example of this approach. At the same time, confirming the ice origin of the Atka diamicts and the data concerning the role of glaciers in their accumulation makes the possibility that diamicts of similar stratigraphic and geographic positions share the same origin more likely. This applies to the stratigraphic counterparts of the Atka Formation in the areas adjacent to the basin of the upper reaches of the Kolyma River: the Gizhiga Formation of the Omolon massif described by Mikhailov *et al.* (1970) and Kashik *et al.* (1990) and the Magan Subformation of the Dulgalakh Formation from the western Verkhoyansk region, which was suggested to be partly of glacial origin by Andrianov and Andrianova (1962). The origin of other assumed ice and glacial deposits from the Upper Paleozoic of northeastern Asia, and especially the products of their reworking, needs serious additional study.

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Sedimentation Peculiarities, Climates, and Geological Events on the Maestrichtian–Middle Eocene Dry Land in Kazakhstan and Central Asia

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Abstract – A wide development of thick weathering crusts (WC) and products of their reworking (humid and arid red beds) on the Maestrichtian–Middle Eocene land platform of Kazakhstan and Central Asia, the similar climate, and other geological phenomena were caused by a peneplain-type paleorelief. The subglobal character of peneplanation processes on platforms of different continents in this time-interval is reflected in sedimentation, paleoclimates, and other geological events.

The Earth's history shows epochs with environments, paleoclimates, and sedimentation that differ greatly from those of the present. The specific character of those epochs was not immediately apparent. S.V. Meyen (1981, p. 43) rightly stated that "climates and the corresponding landscapes, which have no present analogs, could have existed in the past".

A peculiar stage in the Earth's history is, in particular, evident in the Late Cretaceous–Early Paleogene. This will be discussed below, using the continental deposits of Kazakhstan and Central Asia as an example. The region comprises different platform tectonic structures: the Kazakhstan shield, the Altai and Tien Shan epiplatform orogens, and the adjacent West Siberian and Turanian plates. From the Maestrichtian to the beginning of the Middle Eocene, the eastern part of this territory represented a land that was bounded by the Turanian and West Siberian seas to the west and north. Here, in the discussed time interval, peneplanation processes occurred when vertical tectonic movements became stabilized. As a result, two types of landscapes appeared: high denudation plains with small depressions, sometimes with carst funnels (in the area of the Altai, Tien-Shan, and Kazakh shields), and low-lying plains with large depressions (Zaisan, Semipalatinsk, Sarysu, and others).

The Maestrichtian–Middle Eocene continental red and variegated deposits are widespread throughout the territory under consideration; their thickness ranges from 5 - 10 to 100 - 200 m. They rest on the erosional surface of the WC and are overlain by a hiatus and Middle–Upper Miocene sedimentary sequences. The deposits are composed mainly of fine-clastic rocks. In the northern part of the territory, the rocks are kaolinic clays with lenses of quartz sands, often silicified or ferruginized, with boxite deposits in places, whereas in the southern part, the essentially smectitic clayey deposits dominate, which are often gypsinate or carbonate, silicified in places, with inclusions of quatrz sand and silt.

The regional characteristics of these deposits have been given in many publications; general information on their structure, composition, and forming conditions throughout the territory under consideration are presented in monographs (Erofeev and Tsekhovskii, 1982, 1983; Tsekhovskii, 1987). We reconstructed paleoclimates on the basis of the inferences made by paleontologists, which were verified or modified by the results of lithological investigations.

1. PALEOCLIMATE

The climate on the Kazakhstan-Central Asia dry land in the Maestrichtian–Middle Eocene changed through time and space. At that time, based on floral and palynologic data, a warm or hot climate resembling the subtropical dominated in the western part of the region, which was characterized by sea landscapes on the Kazakh shield and adjacent areas of the Turgai Ridge, South Urals, Western Siberia, as well as on the western Tien Shan and Gissar-Turkestan archipelago (Zaklinskaya, 1963; Ponomarenko, 1966, 1971; Pogodaeva and Orlov, 1968; Blyakhova *et al.*, 1987; Makulbekov, 1987a, 1987b; Oleinik, 1987). In the north the climate was humid (sometimes with arid seasons), and areas with hygrophilous evergreen forests as well as open areas resembling humid savanna occurred in the landscapes. In the south (south of 46 - 48°N), the climate became arid, resulting in the appearance of xerophilous vegetation that grew in dry savanna and semidesert landscapes.

In eastern Kazakhstan (the Zaisan depression and its frame), the subtropical climate and associated floras were noted only in the Eocene (Il'inskaya, 1962, 1984; Rzhanikova, 1968; Romanova, 1974, 1982; Vakhromeev and Akhmet'ev, 1977; Akhmet'ev and

Shevyreva, 1989; Akhmetiev, 1985). In the Maestrichtian and Paleocene, the climate became slightly cooler, a fact confirmed by temperate flora elements (the Greenland province) widely distributed in vegetation spectra and by a decrease in the spectra of subtropical flora elements (the Gelinden province). These paleofloral provinces were singled out by Krishtofovich (1955). Flora with prevailing Greenland elements and subordinate Gelinden elements that indicated temperate or transitional climates (from subtropical to temperate) dominated in the latest Senonian through the Danian throughout the eastern periphery of Kazakhstan from the Zaisan to Ili depressions (Erofeev *et al.*, 1972; Makulbekov, 1987a, 1987b). It should be pointed out that from the Maestrichtian to Eocene, hygrophilous vegetation and, correspondingly, a humid climate north of 46 - 48°N were typical of eastern Kazakhstan (similar to its western areas), whereas a more xerophilous vegetation, indicating an arid climate, was characteristic of the area south of this boundary. Thus, both the Eocene subtropical climate and temperate climates of more ancient epochs can be subdivided into humid and arid subtypes. At the same time, some authors inferred (Il'inskaya, 1962, 1984; Akhmet'ev, 1985; Makulbekov, 1987a, 1987b) that along with the general warming of the climate in this territory during the Eocene, which resembles the subtropical everywhere, the climate aridity increases in the second half of the Eocene, which according to Krishtofovich (1955) was an event that occurred on a global level. As a result, the xerophilous features of the subtropical floras of the Poltava province, which replaced the earlier Paleocene vegetation complexes throughout the territory under consideration, became more pronounced.

In addition to the increase in the climate aridity southward, the duration of the arid seasons became longer in the regions distant from the seashore, for instance, in the Zaisan depression, where the xerophilous features of the Maestrichtian–Eocene floras are well pronounced (Makulbekov, 1987a, 1987b).

When characterizing the climate in which the early Cenozoic hygrophilous forests grew, paleobotanists emphasize that, similar to the subtropical, it differed from the latter in a slight seasonal contrast, whereas the paleofloras were characterized by a mixture of elements from different botanical provinces. Therefore, in some regions, such a climate is sometimes distinguished as a separate type, paratropical (Buzek *et al.*, 1990, and others), apparently in comparison to the modern analog of this climate singled out as the forest zone of southeastern Asia (Wolf, 1979), where the average annual temperatures fall between 20 - 25°C. Currently, Akhmet'ev and other Russian botanists use this term when reconstructing the Early Cenozoic climates. Based on the isotopic and magnesium paleothermometry data, the average annual temperature of seas bordering the land under consideration had a range of 20 - 25°C during the Paleocene–Eocene (Ushakov and Yasamanov, 1984; Yasamanov, 1978). In this context, the application of the term "paratropical climate"seems justified. However, the extent to which the ancient paratropical climate corresponds to the modern one – their distinct features, areal distribution, and correlation with other paleoclimates – is still unclear. Therefore, I believe it untimely to single out the paratropical climate in the considered regions without additional research.

Thus, two types of warm climates can be distinguished in the Maestrichtian–Eocene in the territory under consideration: a climate similar to the subtropical one, and a temperate climate, which can possibly be subdivided into humid and arid subtypes. The temperate climates in the Maestrichtian–beginning of the Paleocene and warmer, almost subtropical, climates in the Late Paleocene and Eocene also developed north of the studied territory throughout vast areas of the Asian continent and its northern shelves, i.e., in the areas of the present day temperate, subarctic, and arctic zones (Sinitsyn, 1980; Ushakov and Yasamanov, 1984; Ronov and Balukhovskii, 1981). Nevertheless, special investigations should be carried out to resolve the classification of climates that have become "extinct".

2. SEDIMENTATION

In the latest Cretaceous–earliest Paleogene on the platform land under consideration, vertical tectonic movements become weaker, and the denudation relief levels off, affecting both the Kazakhstan shield uplifts and the areas of the Alpine epiplatform orogeny (Altai, Tien Shan, Djungar Alatau, and others). As a result, denudation peneplains bearing nearly continuous WC appear (Tsekhovskii, 1973, 1987; Erofeev and Tsekhovskii, 1990). They are represented by kaoline or laterite eluvium, several tens of meters thick, in the northern part of the territory in a humid zone, whereas in the southern arid zone a quartz-hydromica-smectite eluvium, 3 - 10 m thick, dominates in places crowned with a calcareous caliche.

It should be pointed out that some geologists dispute the occurrence of intensive chemical weathering of aluminosilicate crystalline rocks under the arid climate, as well as the formation of a clayey eluvial cover, because this process does not presently occur. Nevertheless, it has been established (Erofeev and Tsekhovskii, 1982, 1983) that the chemical weathering increased sharply during ancient epochs of peneplanation not only in the humid zones, but in the arid zones as well. Recently, buried relics of Late Cretaceous–Early Paleogene clayey eluvium were found and described in detail in the Tien Shan region (Erofeev and Tsekhovskii, 1990). In the given eluvium, the weathering (maturing) of initial aluminosilicate rocks ceased at the quartz-smectite stage, and therefore, an erosion of the eluvium and its redeposition facilitated the accumulation of mature quartz-smectite sedimentary rocks.

In the peneplain relief, deep incisions of an erosional system were lacking, and shallow ravines and river valleys did not cut through the eluvial cover, resulting in the reworking of only the WC upper horizons. This caused the involvement of mature material in migra-

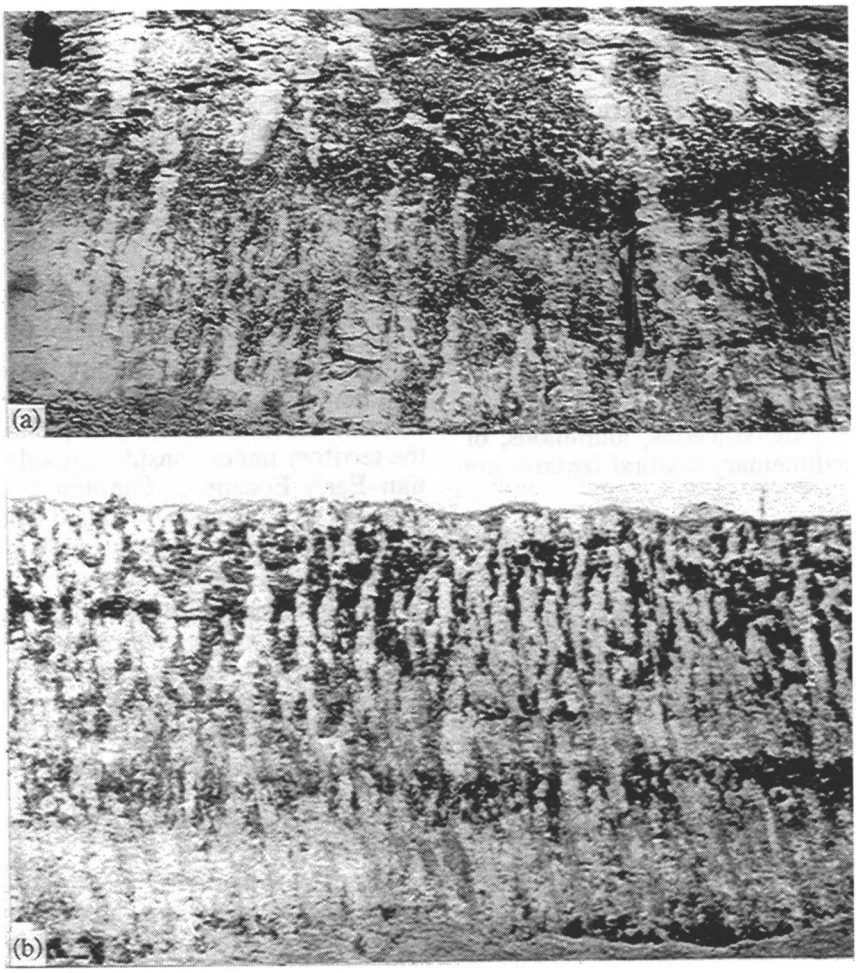

Fig. 1. Maestrichtian–Eocene deposits of the Arkalyk Group, transformed by ferrallitic soils (with vertical axial canals of bleaching) in the sides of open-cuts of the Alkalyk bauxite deposit in Central Kazakhstan.
a – Lateritic soils of the bauxite-bearing ore member, the section is 2.5 m thick; b – ferruginous-siliceous-kaolinic soils of supraore member (the Akzhar Formation), the section is 4.5 m thick.

tion: quartz, caolinite, and lateritic clasts in the humid landscapes, or quartz and smectite in the arid ones.

In the humid landscapes, upon the reworking of red kaolinic or laterite eluvium, facies of plain proluvium, deluvium, small plain rivers that intermittently dried up, lakes, and bogs accumulated under subaerial conditions. Therefore, sequences of red kaolinic clay or quartzose silt with lenses of quartzose sand or sandstone (with goethite-hematite or siliceous-ferruginous matrix) accumulated; sedimentary boxite formed in karst funnels. Under extremely low sedimentation rates characteristic of the peneplain landscapes, the products of WC reworking often undergo repeated weathering after they have been deposited in areas of accumulation. Ferrallitic soils, 0.2 - 0!3 to 3 - 5 m thick, formed over them (Figs. 1a and 1b). As a result, the initial red beds experienced additional weathering, thus attaining a soil texture and turning into mottled red variegated beds, whereas siliceous-clayey, siliceous-ferruginous, ferruginous (goethite-hematite), and, in places, lateritic crusts, often formed in the upper parts of the eluvial profiles.

The lateritic crusts are characterized by numerous bleaching tubular canals formed along the roots of fossil plants (Figs. 1a and 1b) and oriented subvertically, relative to the bedding. The canals often thicken in the upper parts of the soil profiles, branching out and becoming thinner towards their base; the diameter of the canals usually ranges from 0.5 - 1 to 5 - 20 cm, the length from a few dozen of cm to 0.5 - 2.0 m, and their cross section is round or oval. Observations show that a root usually does not fill in the entire canal, but rather occupies its central part. Sometimes the root is well preserved, the result of ferruginization or silicification; but in general, a small hollow remains in its place or it is filled with a porcelainlike clay. The outer part of the tubular canals represents a peculiar tubulated concretion around the root, the composition of which changes into siliceous-kaolinic, kaolinic, or kaolinite-gibbsitic, depending on the soil profile.

The canal sizes allow us to suggest that different vegetations, from forest or shrub to grass, were responsible for their formation.

Relics of ancient soils with lateritic crusts occur at different levels in the Maestrichtian–Lower Eocene. Moreover, the overlying and underlying soil profiles often join in sections and, as a result, the thickness of rock members altered by eluvium may be as much as several tens of meters. As a whole, ancient soil processes were found in 70 - 80% of all the studied deposits. Another type of lateritic crusts related to the cementing of quartz or laterite clastics (with circulation of paleosoil-ground waters) formed in these deposits. Depending on the cement composition, the crusts can be described as ferruginous, siliceous, aluminous, or mixed; their primary sedimentary textural features are well preserved.

Within the ancient arid land in southern Kazakhstan and Central Asia, the reworking products of the ancient, arid WC also accumulated, in general, under subaerial conditions in facies of plain proluvium, deluvium, small intermittent rivers, brackish lakes, solonchaks, and takyrs. These deposits were often altered again by ancient, arid soils (Tsekhovskii *et al.*, 1991), in which carbonate (calcareous or dolomitic, sometimes with palygorskite) caliche and gypsum crusts originated. At the same time, highly mineralized ground water, which migrated through the quatrz sand or silt layers, facilitated the precipitation of gypsum or carbonate, as well as the formation of gypsum or calcareous crusts that cemented the clastic rocks.

Therefore, in both humid and arid ancient peneplanated landscapes, the mature red variegated deposits were formed mainly under subaerial conditions, where facies of temporal flows or small intermittent rivers and lakes dominated. Thin accumulations of sediments (usually from 10 to 100 m), low sedimentation rates (0.4 - 3.6 m/Ma), and numerous intraformational hiatuses fixed by the horizons of thick and mature soils are also characteristic of the area.

Lithological investigations confirm the paleoclimatic reconstructions conducted by paleontologists. The thick kaolinic or lateritic WC and the products of their reworking – quartz-kaolinic and boxite-bearing deposits with relics of intraformational ferrallitic soils and ferruginous, siliceous, and, in places, lateritic crusts – serve as lithologic indicators of a warm (subtropical-tropical) climate in the northern areas. The carbonate-clayey quartz-smectite WC and the products of reworking with similar compositions, usually gypsinate or carbonate, with intraformational soils (resembling the present-day compact or reddish-brown varieties) and calcareous-dolomitic, calcareous, or gypseous caliche and armors, indicate that a warm, arid climate in the southern part of the territory prevailed for a time.

The Maestrichtian–Eocene deposits of the peneplain landscape differ greatly from rock parageneses formed in the dissected plain or mountainous relief, with an intense uplift of the denudation areas and a sinking of the depressions. Grey, terrigenous, often coal-bearing or carbonaceous formations, enriched with coarse clastics and characterized mostly by immature clastic and clayey minerals, usually accumulated under a humid climate, mainly in subaquatic conditions. They are often hundreds or thousands of meters thick, and sedimentation rates reach several tens or hundreds of meters per million years.

3. PECULIARITIES OF SEDIMENTATION PROCESSES

The unusual character of the ancient continental sedimentation resulted from the stabilization of vertical tectonic movements and peneplanation of the relief on the territory under consideration during the Maestrichtian–Early Eocene. A continuous upwarping (epeirogenesis) of all the peneplanated land was accompanied by a slow bending of the peneplain surface and by the formation of a series of intracontinental depressions.

I would like to consider in greater detail the principal environments in which the accumulation of red deposits in humid peneplanated landscapes took place (Fig. 2a) and to compare them with humid environments of contrasting epochs of intensive vertical tectonic movements (Fig. 2b), when ordinary terrigenous or coaly rock parageneses occur in dissected relief environments (for instance, the modern epoch in northwestern Asia).

On paleopeneplains, the accumulation of red deposits started in denudation areas, where only the upper part of the red kaolinic or lateritic WC was subjected to weak erosion (Fig. 2a). No large and deep rivers or lakes could have existed in the accumulative peneplain landscape; they separated into a series of small, shallow, wandering streams and ponds, respectively, where red, mainly fine, clayey, or sandy-clayey sediments accumulated. A peculiar feature of the landscapes was a continuous, mostly clayey cover. In the denudation areas, it was represented by clayey kaolinic areal WC, whereas in the accumulation areas it was essentially composed of clayey products of the eluvium reworking, with small lenses or interlayers of fine-grained clastic rocks.

Such waterproof cover with a thin zone of active water exchange (unlike the modern cover comprised mostly of porous clastic rocks with good collecting properties) prevented the formation of any considerable amount of ground and buried water. The latter, thus, could not feed the ancient rivers, lakes, and bogs. The lack of mountains in the peneplain landscapes deprived the latter of another source of water supply associated with the melting of mountain glaciers. Only one periodical source of water supply remained in these conditions – rain waters. All this affected the humid landscapes and ancient sedimentation.

After a heavy rain, the atmospheric water falling on clayey deposits of the eluvial-sedimentary cover was not able to soak in and be absorbed on a substantial scale. Flowing down to shallow flat depressions and

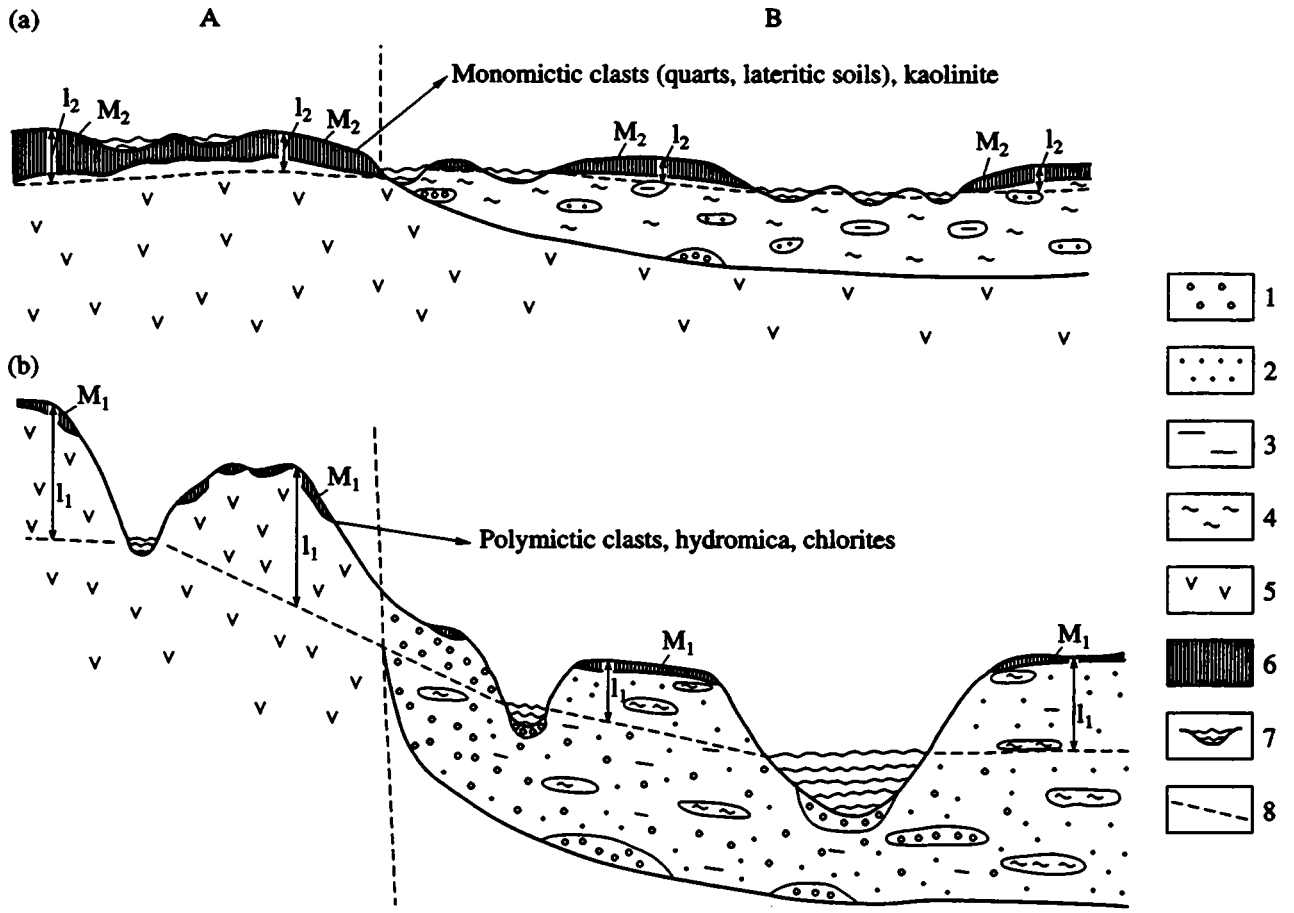

Fig. 2. Humid landscapes of the epochs with a different tectonic regime.
a – a stage of weak vertical tectonic movements and peneplanation of the relief; b – activation stage of vertical tectonic movements, and development of the dissected relief. A – Area of denudation, B – area of sedimentation. 1 – Shingle beds; 2 – sand; 3 – silt; 4 – clay; 5 – host rocks unaffected by weathering; 6 – eluvium cover (weathering crusts and soils); 7 – ponds and streams; 8 – ground water level; l – thickness of the zone of active water exchange ($l_1 > l_2$), M – eluvium cover thickness ($M_1 < M_2$).

small river valleys or lake basins, the water quickly flooded them as well as the vast adjacent plane interfluve areas covered with evergreen forests or the semiopened areas representing a wet savanna; as a result, peculiar periods of flooding occurred. When the rains ceased, the interfluve areas quickly dried and most of the rivers, lakes, and bogs grew shallow or drained without a permanent water supply. As a result, subaerial conditions became predominant, and most of the accumulated deposits were altered by soils. In the given environment, a slope wash of fine terrigenous material by temporary flows from the high to low areas dominated, and a proluvium seems to have accumulated both in the semiopen savanna and in the adjacent forests.

No prerequisite to the burial of organic matter in sediments existed under subaerial diagenesis, so their products of the WC reworking usually maintained the initial red color or became mottled-reddish in old soils. A small amount of organic matter was buried in subaquatic alluvial or lacustrine-bog facies only or a local scale and the stratal-diagenetic bleaching of rocks occurred. It should be emphasized that the present epoch, characterized by the activation of vertical tectonic movements on the continents, is not favourable for the formation of denudation peneplains. Planation surfaces with a thick WC, which developed over vast areas of some continents, are relics of ancient peneplains that are now subjected to destruction (King, 1962). The relics of ancient planation surfaces, which developed on the tops of modern ridges (for instance, on Altai, Saur, Tien Shan and others), are an example. In their turn, the modern vast intracontinental plains (for instance, the East European or West Siberian plains) cannot be attributed to accumulative peneplains. Unlike the considered planation surfaces, they have a more incised drainage system, and occur on the continental blocks highly uplifted over the sea level. This promotes high speed river flow, resulting in the transportation and accumulation of coarser clastics (sand, silt, silty clay, or loam, mostly of an immature composition) in contrast to the peneplain landscapes, where sediments are represented by mature deposits (mainly by clay with subordinate interlayers of silt and sand). The relatively fast downwarping of modern accumula-

tive plains hinders the formation of thick eluvial profiles bearing deep chemical alteration of the original aluminosilicate material, similar to those in the Maestrichtian–Eocene, and decreases the role of soil lithogeneses.

Thus, peneplanation of landscapes elucidates the specific features of ancient humid sedimentation: the prevalence of a subaerial environment, which results in the intense formation of proluvial deposits, an absence of coal-bearing formations, and the widespread development of red deposits, their mature composition, low sedimentation rates, abundance of hiatuses, and others.

The stage considered is distinguished by minimally contrasting paleolandscapes, facies depositional environments, and climates. At that time, a mountain or highly dissected plain relief, and correspondingly the mountain and lowland rivers with hightened hydrodynamics, as well as the large lake basins and boggy areas disappeared throughout vast regions, resulting in the cessation of the accumulation of polymictic or mesomictic deposits and formation of coaly and coal-bearing sequences. Although unstable, the climate remained warm everywhere. The original structures of sedimentary rocks were often concealed by transformations in paleosoils, where uniform superimposed structures of the latter became the determining agents. Therefore, the appearance of the texture, and the composition of the Maestrichtian–Eocene deposits are almost identical everywhere within the considered humid or arid climatic zones.

4. A SPATIAL CORRELATION OF THE STAGE

I pointed out earlier (Tsekhovskii, 1987) that the deceleration stage of vertical tectonic movements and relief planation in the latest Maestrichtian–earliest Paleogene noted in Kazakhstan and adjacent regions affected vast areas of many platforms and was subglobal in character. This is confirmed by the main geomorphological and lithological features of this stage: peneplanated landscapes, thick WC, and the products of their reworking (for instance, red or variegated siliceous-kaolinic or bauxite-bearing sequences on the humid land). Data on the global distribution of the Early Cenozoic planation surface is discussed in the works of King (1962) and his predecessors; some authors emphasize the outburst of the boxite formation in the latest Cretaceous–earliest Paleogene (Bárdossy, 1981; Akaemov *et al.*, 1975; Tsekhovskii, 1987). There is evidence of a wide development of continental siliceous-kaolinic sequences at the Cretaceous–Paleogene boundary within different continents of the Southern Hemisphere (Partridge and Maud, 1990), as well the global character of this phenomenon, which had been inferred by us. The investigations of Gassaway (1985, 1988, 1989, 1990) in North America indicate that continental siliceous-kaolinic sequences are also traced throughout this region. Together with data on the Southern Hemisphere, and the previously discussed data on western Asia, this confirms their subglobal distribution.

Taking northern Asia as an example, it was established (Tsekhovskii, 1987) that the time interval of the considered stage was longest in duration (the Maestrichtian–Middle Eocene) in the inner stable areas of the platform and shortest (the Paleocene–Eocene or a part of the Eocene) in the platform periphery bordering the Cenozoic folding. The latter bears no features of the characterized stage (except for some areas in the Mediterranean region, where lateritic and sedimentary karst bauxites often occur).

Ronov, Khain, and Balukhovskii (1989) also argue in favor of the prevalence of platform land planation in different continents of the Early Cenozoic (Paleocene). They established that the lithological indices of the Maestrichtian–Middle Eocene peneplain landscapes (boxite-bearing, siliceous-kaolinic, or quartz-smectitic sequences related to the WC reworking) show no distinct changes in various regions, remaining similar to those noted in Kazakhstan. This was a result of the common tectonic and climatic conditions that occurred at that time on a global scale: the weakening of vertical tectonic movements and warm climates. The latter are characterized in a series of publications that discuss these questions on a global scale (Ronov and Balukhovskii, 1981; Ushakov and Yasamanov, 1984).

EVENT ANALYSIS AND CONCLUSIONS

The fact that global peneplanation in the latest Cretaceous–Early Paleogene affected vast land areas in different continents in accordance with a global warming allows us to consider some remarkable geological events of that time in a new light (lithological, paleogeobiological, and climatic events).

The lithological events. The wide development of reddish sedimentation on the humid and arid land at the given time interval can be attributed to lithological events. The causes of mass deposition of red formations in the Earth's history are still unsettled. Anatol'eva (1980) and Strakhov (1963) beleive that the warm or hot variable humid and arid climate in the latest Phanerozoic, when vegetation conquered the land, is the main agent defining the sedimentation of red sequences. Sochava (1979) relates this phenomenon to the increase of oxygen content in the atmosphere during various geological epochs and to the growth of its oxidation potential. The data discussed enables us to relate the mass appearance of red formations in humid and arid landscapes of the latest Cretaceous–earliest Paleogene to two main agents: the hot or warm climate and peneplanated landscapes (Tsekhovskii, 1987).

In addition, the domination of the relief peneplanation processes in the time interval considered is confirmed by the data obtained by Ronov and his colleagues (1986) on abnormally low (for the Paleocene) erosion intensity on land and the medium rates of both denudation in continents and sedimentation on land, continental margins, and in oceans. In turn, the

low input of the material derived from the peneplanated land into continental and oceanic sedimentation basins indicates the cause of the hiatuses, which occur in the lower Cenozoic deposits and are fixed in continental sequences not only by erosional surfaces, but also by thick soil horizons.

Paleobiological events. The important role of land relief in producing nutrient resources for seas and oceans has been shown in a series of publications (Tappan, 1968, 1979). They indicate that only a small amount of nutrients are derived from the peneplanated land covered with the WC, which may result in their deficiency in seas and oceans, thus promoting the extinction of biota during that time interval. In fact, the peneplanated Maestrichtian–Middle Eocene land, covered with the WC or its reworked products, was crowned with residual eluvial duricrusts and red caps and highly depleted in main biophile elements vital for organism activity: phosphorus, sulphur, calcium, potassium, and others, including various rare and dispersed elements. This undoubtedly had a negative affect on the organisms in the seas and continents, and, as it is thought, could result in their degradation or extinction when certain levels of the nutrient exhaustion are reached. However, the peneplanated land, which affected sedimentation and biocenoses for a long time (from the Maestrichtian to the first half of the Eocene), appeared to be one but not the main agent causing a change in the organic world at that time. At the same time, the great inpact of this land on the nutrients is evident, and should be taken into account when studying changes in the continental and marine biota.

Paleoclimate. Sinitsyn (1980) considered the general effect of the paleorelief on the paleoclimate. He emphasized that with the lowering of the hypsometric level, the climate on continents should become warmer and more humid due to a decrease in the atmospheric pressure and evaporation. The effect of high mountain ranges on the distribution of rainfall, vertical climatic zonation, and appearance of flooded landscapes in piedmonts is common knowledge. It has been pointed out earlier that the relief peneplanation stage with a warm climate under discussion lacked the arctic, subarctic, and cold temperate belts similar to the modern ones, and only equatorial, subtropical, and warm temperate climatic belts could be singled out with additional subdivision into the humid and arid zones.

In the present work, I do not consider the main causes of global warmings in the Earth's history, which are debatable (Sinitsyn, 1980; Ronov and Balukhovskii, 1981; Ushakov and Yasamanov, 1984). It should be noted, however, that a low hypsometric level of the peneplanated land should benefit warming (Sinitsyn, 1980), and the lack of mountains in the peneplain landscapes resulted in the disappearance of the vertical climatic zonation that contributed to the low degree of climatic contrast. The main reason for the low degree of contrast in the Maestrichtian–Eocene climate lay in the lack of cold and glacier zones in the subpolar parts of the Earth at that time, which prohibited the ingression of cooled air or water with cyclones and currents, respectively, from the arctic areas into the regions under consideration. Therefore, the low-contrast climate can be subdivided into only the warm and warm temperate; distinguishing smaller climatic zones can be done on the basis of the degree of its aridity. In the subpolar regions, the duration of the polar day and night had an additional effect. On the whole, a general regularity was characteristic of the peneplain landscapes: climates were more humid in seaside landscapes and became drier (up to arid) in intracontinental areas. At the same time, peneplain landscapes slightly saturated with water, which we pointed out earlier, the unstable character of the lake-river system, and poor ground-water resources reduced the territories occupied with flooded landscapes and, accordingly, the evaporation intensity in areas of the intracontinental humid lithogenesis and increased the climate dryness in the arid zones, resulting in the expansion of their distribution areas. This also affected the vegetation. The xerophytic features of the Paleocene–Eocene flora in the intracontinental areas (Zaisan depression), which were pointed out earlier, may serve as a point of comparison to the seaside regions (northern Kazakhstan), where it is usually humid (Makulbekov, 1987).

It is noteworthy that although the peneplanation of the vast land areas does not explain the appearance of warm biospheres in the Earth's history, its effect on the character and distribution of climatic zones in these epochs is evident.

In conclusion, it should be emphasized that the epochs of peneplanation of the vast land areas and the warm low-contrast climate also occurred in the older Phanerozoic epochs, apart from the Maestrichtian–Middle Eocene. For instance, in the early Mesozoic, global planation surfaces formed (King, 1962; Gerasimov, 1970), with which the origin of the thick WC as well as an intensive boxite formation in various regions of the Earth are related (Bogatyrev, 1980; Bárdossy, 1977). However, no detailed investigations of this and other epochs of peneplanation have been conducted.

The global peneplanation of the platform land and accompanying synchronous warming of the climates at different stages of the Earth's evolution seem to be the effect of a single cause, which awaits description.

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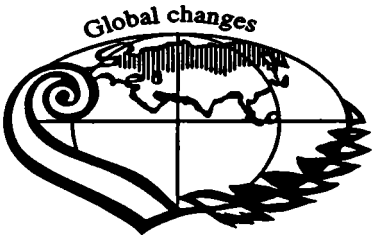

Evolution of the biosphere

Paleogene Episodes of Biogenic Silica Accumulation in the Northern Caucasus and Adjacent Tethyan Regions

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Abstract – The analyzed distribution of siliceous plankton showed that biogenic silica accumulation in a number of Paleogene sections of the northern Caucasus, Crimea, and Mangyshlak was episodic. Seven stratigraphic intervals were assigned, according to when siliceous organisms, i.e., radiolarians, diatoms, silicoflagellates, and sponges, appeared in carbonate-terrigenous deposits to form siliceous sediments (interbeds) under certain facies environments. As was proved by correlation with planktonic foraminiferal and nannofossil zones, the detected levels are synchronous in various sections. Corresponding intervals are distinguishable in the Upper Paleocene (one level), Lower Eocene (two levels), Middle Eocene (two levels), and Oligocene (two levels). Standard biozonation units were recognized in deposits of the Lower-Middle Eocene (the radiolarian *Buryella clinata, Phormocyrtis striata striata*, and *Dyctioprora mongolfieri* zones), in the upper part of the middle Eocene (the silicoflagellate *Dictyocha hexacantha* Zone), and in the Upper Oligocene (the diatom *Bogorovia veniaminii* Zone of tropical scale).

During the Paleogene, the northern Caucasus was a deep part of an epicontinental basin with predominantly carbonate, or carbonate-terrigenous sedimentation in various environmental zones. Despite the highly diverse facies composition of the deposits, the general sedimentation trend was nearly constant throughout the northern Tethyan provinces, where silica accumulation processes were of little importance.

Northward, in a shallower part of the basin located in the southern areas of the Russian platform, mainly terrigenous and siliceous deposits with abundant diatomite and tripoli beds accumulated during the Paleogene. The fact that siliceous organisms widespread in shelf deposits of the northern Tethyan provinces (southern localities of the Russian platform) were absent in adjacent, deeper zones of the same basin seems enigmatic.

Still, available geological descriptions of various localities in the northern Caucasus often report occurrences of "siliceous interbeds", "opokalike clays", and "cherty concretions" in the upper Paleocene, Lower-Middle Eocene, and Oligocene sediments.

Many researchers also reported occurrences of radiolarians and diatoms in the Paleogene deposits of the region, but their paucity, impoverished taxonomic composition, and commonly poor preservation impeded the studies of these organisms, both lithologically and stratigraphically.

Despite this, radiolarian occurrences are definitely known in the upper Paleocene deposits of the Black Sea coastal depression (Chekunov *et al.*, 1976) and the western Caucasus (Borisenko, 1958), in the Eocene sediments of the western Caucasus (Krasheninnikov, 1960; Borisenko, 1960) and Kobystan (Mamedov, 1970), and in the Kuma Formation and its analogs (Borisenko, 1959; Shutskaya, 1970; Moksyakova, 1972). In addition, diatom flora has been reported in the Suzak Formation of Turkmenistan (Teslenko, 1949) and in the Maikop Formation of the western and eastern Caucasus (Grossgeim, 1960; Bogdanovich, 1960; Sheshukova-Poretskaya and Gleser, 1962; Jouse, 1978). This data suggests that there were stratigraphic intervals, when the biogenic silica accumulation was intensive throughout the provinces under consideration, and sometimes dominated carbonate sedimentation.

As a basis for this study, we used a bed by bed analysis of siliceous plankton (radiolarians, diatoms, and silicoflagellates) and some benthic representatives (spicules) from the Paleogene reference sections of the western (Kuban' River), central (Kheu River), and eastern Caucasus (Rubas-Chai River), supplemented with additional data on sections in the Nasypnoe (eastern Crimea), Bol'shoi Zelenchuk, and Urukh localities. N.G. Muzylev and V.N. Ben'yamovskii, respectively, studied nannofossils and foraminifers from the same sections; this simplified our correlation of horizons rich

in siliceous organisms using the Crimea-Caucasus zonation scale based on the distribution of calcareous plankton.

We did not pay much attention to lithology, and studied thin sections only to determine the degree of rock recrystallization in order to infer the primary rock composition and the content of the siliceous organisms.

Several levels of siliceous organism occurrences have been detected in each of the studied reference sections. Despite the paucity of siliceous plankton in many of the discovered assemblages, which hinders detailed stratigraphy, we were still able to correlate them with calcareous plankton zonation and to define which pertinent levels are coeval in various sections. To check the accuracy of our limited data, we compiled all available information on the Paleogene sections of the northern Caucasus, proving that the stratigraphic positions of most of the detected levels are stable throughout the region. Still, the present contribution is so far preliminary, in that the lithology and stratigraphic range of intervals bearing siliceous plankton have yet to be studied in additional localities.

LOWER PALEOCENE

In the eastern Kuban' and central part of the northern Caucasus, the base of the Paleocene consists of silty marls of the El'burgan Formation. In the Kurdzhips River section, Borisenko (1960) discovered radiolarians (mainly Sphaeroidea and Discoidea) associated with foraminifers. In the western Kuban', radiolarians of a similar taxonomic composition occur in marls of the Tsitse Formation, the age analog of the El'burgan Formation (Subbotina, 1950). Unfortunately, there is no data on the exact position of assemblages in the sections and on their correlation to available schemes of calcareous plankton zonation.

UPPER PALEOCENE

The biogenic silica accumulation in the Upper Paleocene deposits is notable everywhere. In western areas (western Kuban' and the flysch zone of the northwestern Caucasus), Lower–Upper Paleocene boundary beds consist of clays and siltstones of the Goryachii Klyuch Formation. In the flysch zone of the northwestern Caucasus, between the Black Sea coast and Pshish River basin, this formation is subdivided into the Shibik, Psekups, Kipyachaya, and Akhtyrskaya subformations, each of them bearing radiolarians associated with agglutinated forms of foraminifers (Borisenko, 1960). In the Glubokaya Balka locality (western Kuban'), the Goryachii Klyuch Formation also contains radiolarians, but in a less preserved state. Only some species of the assemblage, described by Borisenko, occur in the Paleocene deposits of the ocean basins.

The most distinct and easily traced interval bearing siliceous plankton is confined to the upper part of the Paleocene. We studied this interval in the localities of the Kuban' and Kheu rivers.

In the Kuban' River locality, the Upper Paleocene section is represented by the Abaza Formation, which is characterized by alternating dark grey, sometimes micaceous clays and light grey siliceous claystone with occasional concretion segregations. The formation thickness varies from 25 to 30 m. Its siliceous interbeds yielded radiolarian and diatom assemblages.

The poorly to moderately preserved radiolarian species *Cenosphaera* sp., and *Thecosphaera* cf. *larnacium* occur in the lower part of the formation. In its middle part, there is a bed with an abundant ferruginate (?) diatoms species *Coscinodiscus* (?) *vigilans* sensu Gleser. The richest assemblages of diatoms and radiolarians occur in the uppermost part of the Abaza Formation. In addition to smaller Sphaeroidea forms of the *Cenosphaera* and *Thecosphaera* genera, such forms as *Dorcadospyris confluens* Ehrenberg and *Larnocalpis* (?) *smili* Middour were detected. According to Kozlova (1984), the first was encountered in the Syzran' Formation of the Volga provinces, namely in the *Buryella tetradica* Beds, and in the Upper Paleocene–Lower Eocene deposits in the Gulf of Mexico. The second species occurs in the Marsyatka Formation of the eastern Transurals, in the *Cromyocarpus ovatus* Zone (Grigor'eva, 1975) of the west Siberian Talitsa Formation, and in the *Buryella tetradica* Beds of the Volga localities and in the Creek Formation of Missouri. The same beds also contain the diatom assemblage bearing *Melosira sulcata* var. *sibirica*, *Coscinodiscus* aff. *symbirkianus*, *C. decrescens*, and *Pyxidicula weyprechtii*. It is remarkable that the Paleocene diatoms have so far been detected only in sections of the Russian and Siberian platforms, and the Suzak Formation of Middle Asia (Teslenko, 1949). Compared to the Suzak assemblage, the assemblages of other localities under consideration are significantly poorer, but still show elements similar to those observed in the Paleocene diatom assemblages of the Russian platform.

The Abaza Formation age is well substantiated by calcareous plankton of the *Acarinina acarinata* foraminiferal and *Discoaster multiradiatus* nannoplankton zones, both corresponding to the upper Paleocene zones of the Crimea–Caucasus and oceanic scales.

Tracing this level westward, we come to known radiolarian occurrences in the Abaza Formation of the Malyi Zelenchuk locality and in the coeval Ili Formation exposed along the Pshish River (Borisenko, 1960). In addition, we discovered radiolarian fragments and rare diatoms in the upper Paleocene black clays of the Nasypnoe locality, eastern Crimea, from which Muzylev recovered nannoplankton forms characteristic of the upper *Discoaster multiradiatus* Zone. Westernmost counterparts of the described siliceous beds are represented by opokas and opokalike clays of the Mirnenskoe Beds, which have been referred to the Paleocene deposits of the Black Sea coastal depression located west of Evpatoriya (Chekunov *et al.*, 1976).

East of the Kuban' River, a coeval radiolarian assemblage was discovered in the Nal'chik structural zone along the Kheu River. Here, a member of alternat-

ing greenish-grey low-calcareous and light grey opoka-like clays occurs above the Upper Paleocene bed bearing black sapropel with nannofossils of the *Discoaster multiradiatus* Zone, being in turn overlain by massive siliceous claystones, which are olive-grey on a fresh surface, becoming light grey upon weathering. The last member is 35 m thick. In thin sections, the rocks were also classified as recrystallized radiolarites bearing diatoms, though we failed to separate the latter.

In addition to abundant Sphaeroidea and Spongodiscoidea forms, we determined *Cromyodruppa ovoidea, Ommatogramma bilobatus, Spongasteriscus* aff. *cruciferus, Phormocyrtis striata exquisita, Lychnocanoma* sp., *Petalospyris* aff. *foveolata*, and *Dictyomitra* sp. in these radiolarites.

The species *Ph. striata exquisita* and *P. foveolata* appear in the second half of the Paleocene. Some representatives of the *Dictyomitra* genus characteristic of the Upper Cretaceous also occur in the Paleocene; such species as *Ommatogramma bilobatus* and *Cromyodruppa ovoidea* were characteristic of the moderate latitudes of that time (Lipman, 1957; Borisenko, 1959; Kozlova, 1983).

Fig. 1 shows the stratigraphic position of the level described above, and Fig. 2 represents sections where Upper Paleocene radiolarians have been discovered.

As was noted, opokalike rocks and siliceous claystones occur in many sections of the Abaza Formation, but the problem of their primary or secondary silica deposition is still unresolved. As for the sections of our study, the true siliceous beds have only been encountered in the Kheu River section of the Paleocene. Shutskaya (1970) described this interval as an alternation between calcareous, foraminiferal, and siliceous (radiolarian) beds. In the Kuban' River sections of the Abaza Formation, radiolarians and diatoms were only discovered by washing out the samples from the clay interbeds, whereas "siliceous strata" themselves consist of secondary quartz and contain no siliceous organisms. Silica in the Paleocene rocks of the Rubas-Chai section, composed mainly of cherty limestones, is also of secondary origin.

The possible source of secondary silica is an interesting question. The clear stratigraphic position of corresponding beds suggests quite naturally that it originated from dissolving siliceous organisms, yet we have no reliable data to prove this assumption.

LOWER EOCENE

Some Lower Eocene radiolarians were detected in clayey siltstones of the Georgievka Beds (Cherkess Formation) in the Kurdzhipes and Belaya localities (Borisenko, 1960). A more diverse radiolarian assemblage of the *Cenosphaera, Cromyodruppa, Thecosphaera, Ellipsoxiphus, Spongodiscus, Stylodictya*, and *Tricolocapsa* genera was described in the Zybza Formation (Borisenko, 1960). Krasheninnikov (1960) published data on two Lower Eocene radiolarian assemblages from sections of the northern Caucasus located between Krymskaya in the west and the Pshish River in the east. The lower one, occurring in the previously mentioned Zybza Formation, was correlated with the *Globorotalia subbotinae* Zone, while the second assemblage of the Kutaisi Formation was placed on level with the *Globorotalia aragonensis* Zone. Spumellarians of the *Dorylonchidium, Acanthosphaera, Ellipsostylus, Ellipsidium, Staurodiscus*, and *Hagiastrum* genera, which occur widely along with nasselarians in these Lower Eocene assemblages, clearly distinguish them from radiolarians of the Upper Paleocene. Some radiolarian species described by Krasheninnikov (*Lithomespieus* (?) *mendosus, Phormocyrtis turgida*, and *Podocyrtis hirsutus*) are now known to occur in the Lower Eocene deposits of the Pacific and Atlantic oceans, and are used in zonation stratigraphy. In general, the Lower Eocene radiolarians of western Ciscaucasia, having been revised and more accurately classified according to the modern taxonomic system, seem useful for distant correlations.

Our own collection of fossils is somewhat poor. We discovered an assemblage of smaller Sphaeroidea forms associated with diatoms (*Pseudopodosira* sp.) in the light grey clays of the Kuban' River locality near Cherkassk. According to coexisting foraminifera, the host deposits are referred to the lower horizons of the *Globorotalia subbotinae* Zone (determinations by Ben'yamovskii), which is consistent with the stratigraphic position of the lower assemblage distinguished by Krasheninnikov.

Radiolarians were also detected in thin sections of calcareous sandstones from the Rubas-Chai locality. According to nannoplankton, these rocks correspond in age to the *Marthasterites tribrachiatus* Zone (NP12), which is coeval with the *Globorotalia aragonensis* foraminiferal zone. Therefore, this level correlates to the second lower Eocene zone of Krasheninnikov's subdivisions.

MIDDLE EOCENE

In the western Kuban' localities, radiolarian assemblages are known to occur in the Kaluga and Khadyzhensk formations (Borisenko, 1959). At present, both formations can be correlated to the *Acarinina bulbrooki* foraminiferal zone, and *Nannotetrina fulgens* Zone on the nannoplankton scale. The Kaluga Formation is composed of green clays and siltstones grading upward into more calcareous varieties. The Khadyzhensk Formation consists mainly of green marls. In the Pshish locality, it contains interlayers of light lilac waxlike kills (Grossgeim, 1960). Borisenko (1959) described *Cenosphaera reguris, Thecosphaera rotunda*, and *Ellipsidium cultum* forms as characteristic of the Kaluga Formation, and the less diverse radiolarian assemblage as being from the Khadyzhensk Formation. It is noteworthy that radiolarians occur at this level in most sections of the western Caucasus (Grossgeim, 1960).

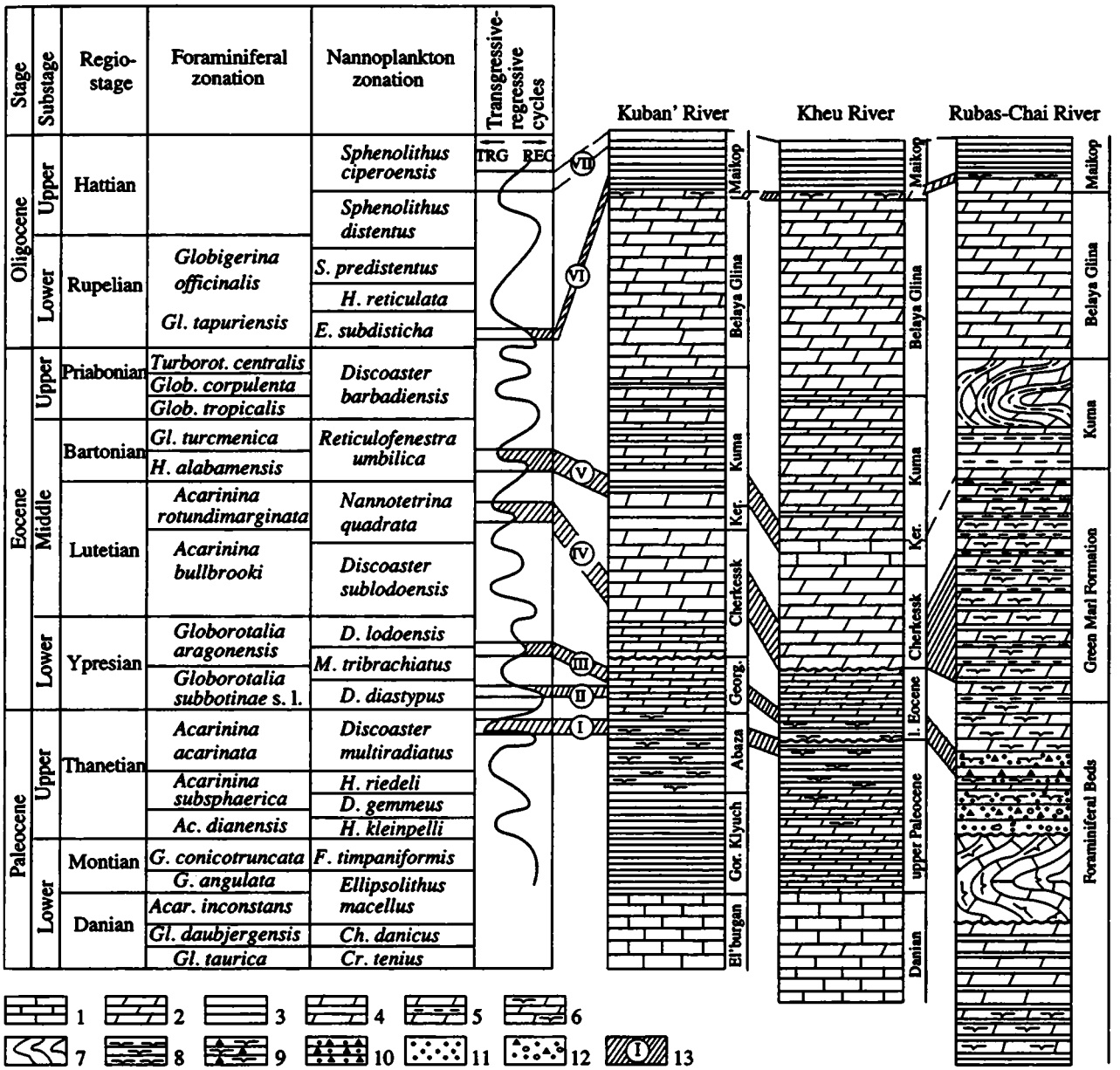

Fig. 1. Correlative occurrence levels of siliceous plankton in the reference sections of the western, central, and eastern Caucasus.
1 – Limestones; 2 – marls; 3 – clays; 4 – bituminous marls; 5 – claystones; 6 – siliceous marls; 7 – syngenetic landslide deposits; 8 – siliceous claystones; 9 – the same with glauconite; 10 – glauconitic sand- and claystones; 11 – glauconitic sand- and gravelstone; 12 – occurrence levels of siliceous plankton; 13 – levels of biogenic silica accumulation.

We separated abundant radiolarians from green marls of the Kheu River locality. According to the nannoplankton finds, the rocks correspond to the middle part of the *Nannotetrina fulgens* Zone. Specimens of *Spongodiscus* sp., *Pendrospyris* sp., *Haliomma* sp., *Cromyodruppa* aff. *ovoidea*, and some others have been distinguished among recrystallized radiolarians. Associated diatom forms include *Pyxidicula johnsoni*, *P. edita*, and *Coscinodiscus decrescenoides*.

The richest assemblages of siliceous organisms were recovered in the Rubas-Chai locality of Dagestan. The section consists of hard siliceous sandstones poor in calcareous components, alternating with softer green claystones (Fig. 1). The lower member of the section (15 m thick) is represented by claystones with thin interlayers of sandstones. The overlying member of massive cherty sandstones is 5 m thick and separates the lower unit from the upper member, which has a clear flyschoid structure and is 25 - 30 m thick. Sandstone beds become thicker towards the top of the section.

Claystones, colored bright greenish-blue in the lower part and green elsewhere, are the most interesting rocks of the section. In thin sections they appeared to be spongolites, or spongo-radiolarites sometimes bearing

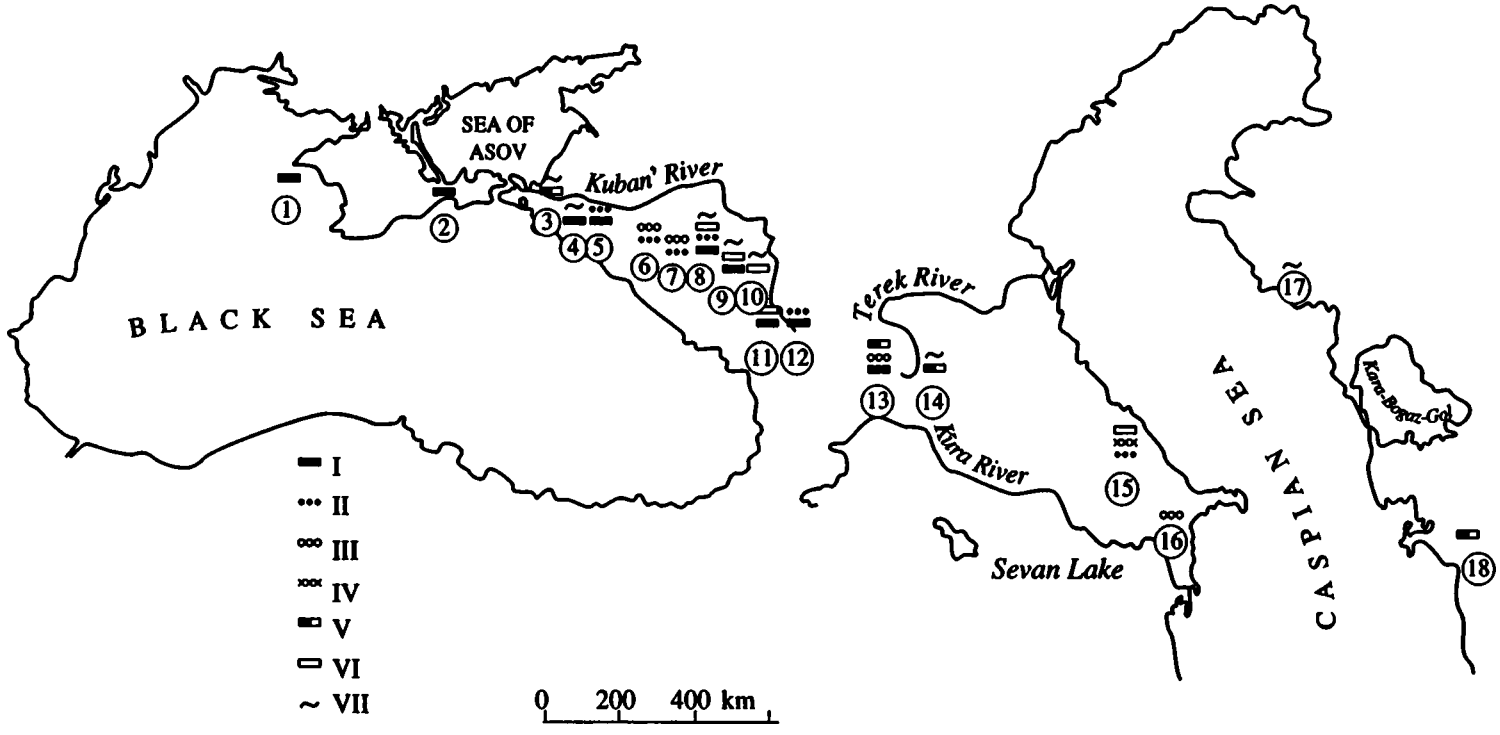

Fig. 2. Geographic positions of sections bearing siliceous plankton.

Localities: 1 – Evpatoria (Chekunov *et al.*, 1976); 2 – Nasypnoe (our data); 3 – Anapa (Bogdanovich, 1960); 4 – Shibik River (Grossgeim, 1960; Sheshukova-Poretskaya and Gleser, 1962); 5 – Glubokii Yar; 6 – Balka Kipyach'ya; 7 – Balka Glubokaya; 8 – Pshish River (our data); 9 – Kurdzhips River (Grossgeim, 1960; Krasheninnikov, 1960; Bogdanovich, 1960); 10 – Laba River (Bogdanovich, 1960); 11 – Belaya River (Grossgeim, 1960); 12 – Kuban' River (our data); 13 – Kheu River (our data); 14 – Urukh River (Shutskaya, 1970; Bogdanovich, 1960); 15 – Rubas-Chai River (our data); 16 – northern Kobystan (our data); 17 – Karagie Depression (Jouse, 1978); 18 – Bol'shoi Balkhan (Shutskaya, 1970). Occurrence levels of siliceous plankton: I – Upper Paleocene; II, III – Lower Eocene; IV, V – Middle Eocene; VI, VII – Lower-Middle Oligocene.

diatoms in addition to sponge spicules and radiolarians. Despite the recrystallization, some of these beds yielded moderately and well preserved radiolarians. Their species in the lower claystone member include *Phormocyrtis striata exquisita*, *Ph. striata striata*, *Ph.* cf. *reticula*, *Haliomma californica*, *H. melitomma*, *H.* aff. *senta*, *Entapium regulare*, *Stylosphaera irinia*, *St. coronata coronata*, *Lychnocanium* aff. *concinnum*, *Lophophaena macrencephala*, *Calocycloma ampulla*, *Theocampe mongolfieri*, *Hexaspyris fiscella*, *Lithochytris archaea*, and *Lophoconus* sp.

The taxonomic composition of this assemblage definitely corresponds to the standard *Dyctioprora mongolfieri* Zone of the tropical radiolarian scale. This is evident from the fact that the index species coexist with *Calocycloma ampulla*, *Lithochytris archaea*, and *Podocyrtis striata striata*, which are characteristic of oceanic deposits of that stratigraphic interval. Some species of the assemblage do not occur in the Paleogene sediments of the tropical and subtropical provinces, but are well known in the west Siberian and Volga regions (Kozlova, 1984; Kozlova and Gorbovets, 1966), e.g., *Stylosphaera irinia*, *Hexaspyris fiscella*, *Haliomma senta*, *H. melitomma*, *Lophophaena macrencephala*, and some others. Muzylev's data on nannoplankton indicates that the deposits under consideration correspond in age to the *Nannotetrina fulgens* (NP15) Zone.

The upper claystone member yielded a rich assemblage differing slightly in composition. Its basic species are: *Lychnocanoma bellum*, *Sethocyrtis* aff. *principi*, *S. babylonis*, *S. bicamerata*, *Clathrocyclas universa*, *Porodiscus harlestonensis*, *Druppatractus* aff. *tennus*, *Prunophyle affinis*, *Stylosphaera coronata coronata*, *St. coronata laevis*, *Lithelius foremanae*, and *Dictyoprora mongolfieri*. Accessory forms include: *Thecosphaera californica*, *Sethocyrtis mitrus*, *Spongodiscus radiatus*, *Coniforma antiochensis*, *Periphaena heliasteriscus*, and *Haliomma melitomma*.

In general, this assemblage is impoverished in tropical and subtropical species (*Lychnocanoma bellum*, *Theocampe mongolfieri*, *Periphaena heliasteriscus*, and *Stylosphaera coronata laevis*). It seems relatively younger than the assemblage of the *Dictyoprora mongolfieri* Zone, due to the presence of *Periphaena heliasteriscus* and *Thecosphaera californica* species more typical of the second half of the middle Eocene.

It should be noted that in all Lutetian deposits bearing siliceous organisms (Fig. 2), they are confined to sediments designated as a "formation of greenish-grey clays and blue marls" in western Ciscaucasia (Bogdanovich, 1960), or as a "formation of green marls and limestones" in eastern Ciscaucasia (Golubyatnikov, 1947). In most sections, these green-colored deposits are attributed to the *Nannotetrina fulgens* (NP15) Zone on the nannoplankton scale. Their color is caused by the finely dispersed glauconite, while the rock composition varies from pure carbonate in central Ciscaucasia to carbonate–volcanogenic in the western province and terrigeneous–siliceous in the eastern province. We studied rocks of the latter in the Rubas-Chai locality of Dagestan. Northward, in the Dzhinabi-Chai (Ullu-Chai basin) and Khalargonk river valleys, the same rocks were discovered by Boiko *et al.* (personal communication).

The next occurrence level of siliceous organisms is related to the Kuma Formation. Radiolarians of this level have been recorded in sections of the Shikik, Zybza, Ili, and Khabdi rivers located in western Ciscaucasia (Grossgeim, 1960). However, the most complete list of species was reported from the Glubokii Yar locality, where Borisenko determined such forms as *Cenosphaera harskaya*, *Dorilonchidium venustus*, *Staurolondra ifidelia*, *Cenellipsus fusiformis*, *Sethocyrtis bacaformis*, and *Theocorys guleatus*.

In central Ciscaucasia, we distinguished this level in the Kheu River section. Here, lower bentonitic clay interbeds of the Kuma Formation yielded radiolarian and diatom assemblages. Near the base interval, they contain radiolarians of *Cromyodruppa*, *Cenosphaera*, and *Spongodiscus* genera, and somewhat higher we detected *Theocotyle* sp., *Clathrocyclus* sp., *Lychnocanium* sp., and *Canocaryomma lentis*. The last two species occur in the Tasaran Formation deposits of the North Aral province and Turgai depression, which correlates them to the Kuma Formation of Ciscaucasia (Moksyakova, 1972). These species also occur in the Kuma deposits in the Turan plate (Fig. 2).

The coeval diatom assemblage includes: *Paralia camaruensis*, *Melosira sulcata* var. *sibirica*, *M. fausta*, *M. architecturalis*, *Pseudopodosira westii*, *Coscinodiscus obscurus* var. *concavus*, *Pyxidicula enermis*, and *Triceratium ventriculosa*, which are characteristic of the Eocene deposits in southern provinces of the Russian platform.

It should be noted that silicoflagellates of this level include specimens of *Dictyocha hexacantha*, which serves as the index species of Martini's zonation scheme for silicoflagellates characterizing the middle and basal Upper Eocene deposits.

Fragmented diatoms have also been detected in transitional beds between the Kuma and Belaya Glina formations.

At the base of the Kuma marls bearing siliceous fossils, we also discovered planktonic foraminifers, the assemblage of which is intermediate in composition as compared with species of the *Hantkenina alabamensis* and *Globigerina turkmenica* zones. The top layers of the Kuma Formation yielded foraminifers that correlate them to the upper part of the *Globigerina turkmenica* Zone (determinations by Ben'yamovskii).

OLIGOCENE

In the western Kuban' provinces (Kuban' and Fars rivers), pyritizated radiolarians associated with abundant planktonic and benthic foraminifera, spore and pollen, and fish and plant remnants occur in grey calcareous clays (Khadum Horizon) of the Maikop Group (Bogdanovich, 1960).

Throughout the western Ciscaucasia (from Anapa to the Pshish, Pshekha, Belaya, and Laba rivers), the Khadum Horizon bears sponge spicules and scarce radiolarians that are sometimes accompanied by diatoms. Bogdanovich reported, however, only their large benthic forms (*Isthmia dzaboi* and *Pseudopyxilla* sp.) associated with foraminifers and radiolarians, because special diatom separation methods have not been used.

In the Rubas-Chai locality of eastern Ciscaucasia, we discovered abundant and diverse radiolarians at the base of the Maikop deposits, where due to recrystallization, they are poorly to moderately preserved. Such forms as *Lithocampe* sp., *B. porodiscus* sp., *Dictyoprora pirum*, *Carpocanopsis* sp., *Cryptoprora* sp., *Hexacontium micrum*, *Theocotyle* sp., and *Lychnocanoma* aff. *trifolium* occur most commonly. Species of *Lithocampe* and *Carpocanopsis* genera along with specimens of *Dictyoprora pirum* and *Lychnocanoma* aff. *trifolium* indirectly indicate the Early Oligocene age of the host deposits. In tropical and subtropical oceanic provinces, specimens of *Lithocampe* sp. occur from the Oligocene to the early Miocene, species of the *Carpocanopsis* genus, since the Late Eocene, the *Dictyoprora pirum* representatives, in the Late Eocene to the Oligocene, and *Lychnocanoma trifolium* is an Oligocene form. Muzylev concluded that this stratigraphic level can be correlated with the *Ericsonia subdisticha* Zone (NP21) on the nannoplankton zonation scheme.

Higher horizons of the Maikop Group have not been subdivided according to units of calcareous plankton zonation, and therefore the stratigraphic study and correlation of their levels using siliceous plankton is a very important goal.

The second Oligocene level of silica accumulation is in the middle part of the Maikop Group. It was studied in detail in the northern Karagie depression of southern Mangyshlak, where diatomites compose the upper, 20-m-thick member of the Karagie Formation, the lower part of which consists of "fish shales." Jouse (1978) described a rich diatom assemblage of *Synedra jouseana* f. *linearis*, *Hemiaulus polysystinorum*, *Pseudotriceratium chenevierii*, and *Hyalodiscus radiatus*, which define the Late Oligocene–Early Miocene age of the host diatomites. Coexisting silicoflagellates *Corbisema archangelskiana* and *Mesocena apiculata inflata* restrict their age to the Late Oligocene, while available specimens of *Kozloviella* aff. *edita* of a narrow stratigraphic range allow us to correlate the level under discussion with the NP24 - NP25 zones on the nannoplankton scale. Jouse referred this assemblage to the *Coscinodiscus vigilans* Zone, the Late Oligocene unit of her zonation scheme for oceanic plankton. At present, this zone can be correlated with the *Bogorovia veniaminii* Zone of the Late Oligocene, as it is defined in the standard tropical biozonation (Fenner, 1985).

In the northern Caucasus, the Assa Formation of the Maikop Group is, in its stratigraphic level, similar to the Karagie Formation. In the Urukh River locality of the Nal'chik structural zone, upper clays of this formation overlying "fish shales" contain diatoms of the *Isthmia* genus along with radiolarians and sponge spicules (Bogdanovich, 1960). Westward, in the Adygeya uplift, the formation was thought to correlate to the Karadzhalga and Gora Vysokaya formations. Rocks of the latter yielded radiolarians of *Cenosphaera*, *Stylodyctia*, *Spongodiscus*, and *Cenodiscus* genera, diatoms of *Isthmia* genus, and sponge spicules. To the west of Khadyzhensk, the thickness of the Maikop Group decreases to 20 - 30 m, and a spongolite horizon up to 2 - 3 m thick is located in its middle part.

First Jouse, and then Sheshukova-Poretskaya and Gleser (1962) analyzed diatoms from the middle part of the Maikop Group in the Shibik River locality. These researchers described an assemblage of siliceous organisms from diatomites, which included more than 53 species of diatoms, silicoflagellates, and ebridians, with specimens of *Craspedodiscus elegans*, *Thalassiosira excentrica*, *Hemiaulus* sp., *Synedra jouseana*, and *Dictyocha staurodon* among them. As a whole, this assemblage correlates well to that from the Karagie Formation of Mangyshlak, but does not contain reference oceanic species solely characteristic of the Upper Oligocene.

CONCLUSION

The Paleogene deposits of the northern Caucasus province and adjacent areas show several stratigraphic levels, where siliceous microorganisms dominate over calcareous plankton, and sometimes serve as rock forming components.

I. The oldest level is traced within the Upper Paleocene, and corresponds to upper parts of the nannoplankton *Discoaster multiradiatus* and foraminiferal *Acarinina acarinata* zones. In some sections, radiolarite interlayers occur.

II, III. Two levels of siliceous plankton have been distinguished in the Lower Eocene. Radiolarian assemblages from these stratigraphic levels contain species characteristic of the *Buryella clinata* and *Phormocyrtis striata striata* zones of the standard zonation scheme for oceanic radiolarians, and thus correlate to the *Globorotalia subbotinae* and *Globorotalia aragonensis* foraminiferal zones.

IV, V. Two levels of biogenic silica accumulation can also be traced in the Middle Eocene. In eastern Ciscaucasia, the first is characterized by the deposition of siliceous, remarkably thick sediments such as spongolites, tripolis, and others. The rocks contain two radiolarian assemblages, among which the lower one is referred to the standard *Dictyoprora mongolfieri* Zone. The richer, upper assemblage cannot be reliably correlated with any standard radiolarian zone. Its range seems to correspond to the largest part of the Lutetian, and correlates to the *Nannotetrina fulgens* Zone of nannoplankton zonation.

The second level of biogenic silica accumulation is related to the Kuma Formation, which bears diatoms, silicoflagellates, and radiolarians in its basal deposits. The silicoflagellate assemblage corresponds to that of the *Dictyocha hexacantha* Zone. The level is referred to

the Bartonian, and can be correlated to the *Globorotalia turkmenica* Zone of foraminiferal scale.

The Upper Eocene deposits are barren on the siliceous organisms.

VI, VII. The Oligocene interval also comprises two levels of biogenic silica accumulation. The lower one, located at the base of Khadum Formation, is characterized by occurrences of radiolarians, diatoms, and sponge spicules, and corresponds to the nannoplankton *Ericsonia subdistincha* Zone. The upper level is represented in some localities by diatomite beds bearing a diatom assemblage resembling that found in the *Bogorovia veniaminii* sensu Fenner Zone. The latter corresponds to the upper part of the Oligocene, and correlates to the nannoplankton *Sphenolithus distentus–Sphenolithus ciperoensis* zones.

Thus, the standard zonation units of the oceanic radiolarian scale can be traced only in siliceous plankton assemblages of the lower and basal Middle Eocene, while diatom and silicoflagellate zonation is possible only for the terminal middle Eocene and upper Oligocene assemblages.

The correlation of biogenic silica accumulation levels with zonation units of calcareous plankton enables us to estimate their position on Musylev's curve (in press),which depicts transgressive-regressive cycles in the southern provinces of the CIS (Fig. 1). As one can see, episodes of biogenic silica accumulation were initiated, as a rule, at the beginning of transgressive stages, or during transitions from the transgessive to regressive cycle phases.

Reviewers N.G. Muzylev and Yu.B. Gladenkov

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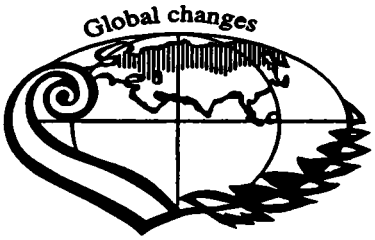

Evolution of the biosphere

Biota of Transitional Pliocene–Pleistocene Sediments in the Kamchatka Region (Diatoms, Mollusks, and Foraminifers)

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Abstract – Faunal (mollusks and foraminifers) and floral (diatoms) assemblages from the Pliocene–Pleistocene beds of eastern Kamchatka are characterized. Three benthonic and planktonic complexes were distinguished in the interval from 2 to 0.5 Ma. The correlation between diatom zones and the paleomagnetic scale provided age determinations. The paleoenvironment in the North Pacific during the Late Pliocene–Early Pleistocene, i.e. the formation time of the recent marine biota, was characterized.

INTRODUCTION

The detailed study of Pliocene geologic events in different parts of the paleobiosphere is a major task in modern geology. Such a study may significantly help both to refine methods of deciphering more ancient geologic events, and to forecast natural environments in the coming centuries.

During the course of our investigations, researchers of the Geological Institute, Russian Academy of Sciences, studied the Pliocene sequences in Kamchatka, Sakhalin, Chukotka, Arctic regions, Iceland, and others, over a period of years in order to better understand and correlate Pliocene geologic events in the Holarctic province. Some materials concerning these topics have already been published (Gladenkov, 1978; Gladenkov *et al.*, 1980; Petrov, 1982; Basilyan *et al.*, 1991). A large monograph on the Neogene of eastern Kamchatka was recently published (Gladenkov *et al.*, 1992). These works deal with many problems concerning Pliocene stratigraphy and paleogeography, and, among them, the history of the ancient Bering Strait. However, there is insufficient data on the transitional Pliocene–Quaternary interval. This data is of particular importance because it allows us to judge the peculiarities of recent biota formation and the events of the terminal Pliocene and initial Pleistocene.

Unfortunately, sections bearing transitional Pliocene–Pleistocene beds are extremely rare in the North Pacific. Thus, detecting them in eastern Kamchatka is rather important, even more so as they yielded numerous paleontologic remains.

Within eastern Kamchatka, the Pliocene–Quaternary marine deposits are only locally distributed. They are represented mainly by siltstones, sandstones, and conglomerates that accumulated on the open shelf and the continental slope. They are distinguished as the Ol'khovaya Formation, which is most completely characterized, in paleontological terms, on the Kamchatskii Peninsula (Fig. 1).

The Ol'khovaya Formation was identified by O.S. Vyalov in 1940. In the 1960s, it was subdivided into two subformations (Bespalyi *et al.*, 1972). Later, in the 1970s, the deposits were studied by G.N. Borzunova, Yu.B. Gladenkov, O.M. Petrov, I.M. Khoreva, M.N. Khotin, and others. Petrov (1982) described the formation and published a monograph on the mollusks. Foraminifers were studied by Bylinskaya and Khoreva (1985), and diatoms, by Dolmatova (1980). Paleomagnetic research was done by Pevzner (Petrov, 1982). At the end of the 1980s, additional investigations by A.E. Basilyan, K.B. Barinov, and A.Yu. Gladenkov provided new data, which served as the basis for this paper.

LITHOLOGY

Deposits of the Ol'khovaya Formation unconformably overlie deformed rocks of the Cretaceous and Paleogene and compose gentle folds that experienced faulting. According to lithologic composition, the Ol'khovaya Formation is subdivided into two subformations (Fig. 2).

The lower Ol'khovaya Subformation is exposed in shore scarps and at water edges of the Mutnaya and other rivers located on the southwestern Kamchatskii Peninsula. Seven members were distinguished within it (upward from the base):

(1) Interlayering of loose yellow-grey and grey conglomerates, gravelstones, coarse- and medium-grained sandstones. Conglomerates, sometimes with boulders up to 0.4 m in size, and gravelstones predominate in the lower part, whereas the upper one is composed of bedded sandstones and loose sands containing layers of different conglomerates (0.08 - 0.5 m thick) with ripple marks, leaf imprints, and *Mytilus* shell fragments. At the top of the sequence, the interlayering of sandstones with gravel and rare dispersed pebbles up to 10 m thick occurs. They yield shell fragments of the mollusks *Astarte olchovica* Petrov, *Cyrtodaria kurriana* Dunker, *Liocyma fluctuosa* (Gould), *Spisula voyi* (Gabb), and others. The total thickness is 110 - 120 m.

(2) Blue-grey silty fine-grained sandstones, indistinctly bedded, with splinter and imbrication cleavage. They commonly contain dispersed pebbles, sometimes accumulated in lenslike beds, as well as rounded (up to 0.3 m) and lenslike (1.2 × 0.4 m) dispersed carbonate concretions. In the Mutnaya-A section, this member contains light-brown volcanic ash layers up to 0.04 m thick. The deposits yield shells of *Cryptonatica clausa* (Broderip et Sowerby), *Aforia circinata* (Dall), *Musculus niger* (Gray), *C. kurriana* Dunker, *Hiatella arctica* (Linne), *Ciliatocardium ciliatum* (Fabricius), *Macoma calcarea* (Gmelin), and others. The total thickness is 40 - 120 m.

(3) Massive blue-grey sandy siltstones with rare dispersed pebbles, gravel, and rare lenslike (0.08 - 0.15 m) carbonate concretions. Beds of clayey dark-grey siltstones with a H_2S odor, up to 0.15 m thick, occur. *Volutopsis planus* Petrov, *Boreotrophon pacificus* (Dasil), *Quasisipho torquatus* Petrov, *Nuculana lamellosa lamellosa* (Leach), *Yoldia toporoki* Scarlato, *Musculus niger* (Gray), *Ciliatocardium ciliatum* (Fabricius), and other mollusks are found there. The total thickness is 35 - 60 m.

(4) Dark blue-grey dense massive siltstones, indistinctly bedded, with coarse splinter cleavage, rare small dispersed gravel, and shells of *Megayoldia olchovica* Petrov, *M. thraciaeformis* (Storeer), and others. At the foot of Vysokaya Mountain, sandy siltstones with large lenses of conglomerate yield *Calyptogena lachtakensis* Petrov, *Thyasira* sp., and others. The total thickness is 110 m.

(5) Indistinctly bedded sandy siltstones with splinter cleavage, common dispersed pebbles, and rare rounded carbonate (up to 0.15 m) concretions. In the lower part, lenslike beds of gravel sandstones (0.02 - 0.4 m) occur, whereas at the base there is a sandy to silty gravelstone layer 2.0 m thick. The following mollusks were found in the member: *Leionucula tenuis expansa* (Reeve), *Nuculana lamellosa lamellosa* (Leach), *Yoldia toporoki* Scarlato, *Musculus niger* (Gray), *Ciliatocardium ciliatum* (Fabricius), etc. The total thickness is 35 - 40 m.

(6) Grey-blue greenish silty sandstones, indistinctly bedded, with a clotted cleavage, common dispersed pebbles, and shells of *Yoldia toporoki* Scarlato, *Musculus niger* (Gray), *Ciliatocardium ciliatum* (Fabricius), and others. In the upper part, at 30 m, a volcanic ash layer 0.03 m thick was found. The total thickness is 35 m.

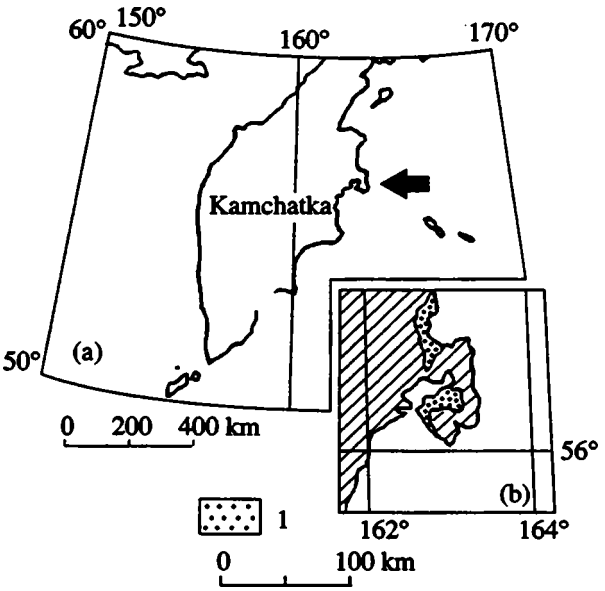

Fig. 1. Distribution of the Ol'khovaya Formation.
a – Position of the Kamchatskii Peninsula; b – distribution areas of the Ol'khovaya Formation on the Kamchatskii Penisula; 1 – Ol'khovaya Formation.

(7) A lenslike cross-bedding of coarse-grained greenish-grey sandstones (0.2 - 0.4 m) and loose dark-grey gravelstones (0.1 - 0.3 m) containing a variety of loose conglomerate lenses (up to 0.4 × 1.5 m). The deposits yield single specimens of *Cyrtodaria kurriana* Dunker. The total thickness is approximately 30 m.

The thickness of members varies along the strike. The total thickness of the subformation exceeds 500 m.

The lithologic composition of the upper subformation also varies significantly along the strike. In the southwestern part of the peninsula, its deposits unconformably overlie the lower Ol'khovaya Subformation and are represented by a thick (730 m) sequence of cross-bedded conglomerates, gravelstones, and sandstones, whereas in the central part, it consists of sandy siltstones and sandstones (approximately 200 m thick). Fossil remnants occur infrequently, mainly in the middle part of the subformation, in the Ol'khovaya-1 River section situated in the central area of the peninsula. The following mollusks were found there: *Cryptonatica clausa* (Broderip et Sowerby), *Trichotropus insignis* Middendorff, *Nuculana minuta sachalinica* Scarlato, *Chlamys beringianus* (Middendorff), *Astarte rollandi* Bernardii, *Cyclocardia ventricosa* (Gould), and others.

The occurrence of numerous dispersed, possibly ice-rafted pebbles in deposits of the formation is noteworthy. At the same time, they contain a great amount of pyroclastic material, mainly acidic in composition.

Various groups of organic remains were described in the Ol'khovaya Formation, i.e., mollusks (identified by A.E. Basilyan), foraminifers (M.E. Bylinskaya), and

diatoms (A.Yu. Gladenkov). A brief description of them is given below.

MOLLUSKS

One hundred and fourteen forms of bivalve mollusks and gastropods were found in the Ol'khovaya Formation (refer to the table). We distinguished two mollusk assemblages that characterize faunal beds. The *Quasisipho torquatus–Yoldia* (*Cnesterium*) *toporoki* assemblage was identified in the lower Ol'khovaya Subformation (marked by an L in the table). It contains 104 forms (88 species). The upper Ol'khovaya Subformation (marked by a U in the table) yields the *Nuculana minuta sachalinica–Chlamys* (*Chlamys*) *beringianus* assemblage, which consists of 24 species. Its composition had, in general, common features inherited from the early Ol'khovaya fauna.

Recent species (85 - 90%) are the basis of the early Ol'khovaya assemblage. Boreal-arctic forms constitute more than 50% of the total assemblage, whereas of the widespread boreal species, more than 20% are characteristic of the sublittoral and upper bathyal zones in the boreal Pacific. The *Quasisipho torquatus–Yoldia* (*Cnesterium*) *toporoki* assemblage is marked by a cooccurrence of arctic, high-boreal forms *Leionucula inflata romboides* Scarlato, *Nuculana lamellosa lamellosa*, *Astarte* (*Nicania*) *montagui fabula* (Reeve), *Cylichna occulta* (Mighels), and low-boreal species *Lunatia pila* (Pilsbry), *Yoldia* (*Cnesterium*) *toporoki*, and *Cyclocardia erimoensis* (Tiba). This indicates the relatively ancient age of the assemblage and an incomplete correspondance to recent malacofauna.

The upper Ol'khovaya Subformation yields a poorer mollusk assemblage. This *Nuculana minuta sachalinica–Chlamys* (*Chlamys*) *beringianus* assemblage contained three extinct species. Biogeographically, it is similar to the early Ol'khovaya assemblage, but lacks such arctic representatives as *Nuculana lamellosa lamellosa*, *Astarte borealis placenta*, and subspecies of *Astarte montagui*. Thus, it is similar to recent sublittoral fauna of the boreal Pacific.

FORAMINIFERS

Deposits of the Ol'khovaya Formation yield 70 species of calcareous benthonic and planktonic foraminifers. We distinguished three assemblages characterizing the faunal beds, which differ in their share of extinct forms and some peculiarities in their taxonomic composition. In all of the assemblages, Elphidiidae, Islandiellidae, and *Buccella* dominate. The samples contain from several hundreds to 2 - 3 thousand specimens per 50 g of dry sediment. The number of extinct species decreases upward from the base of the formation.

Basal sediments of the lower Ol'khovaya Subformation, which were distinguished as the *Buccella conica–Islandiella excavata* Beds, yield the following 8 (about 18%) extinct species: *Pseudopolymorphina lautenschlaegerae* Kusina, *Buccella conica* Volosh., *B. niigataensis* (Husez. et Maruh.), *Retroelphidium hughesi* (Cush. et Grant), *Islandiella sulcata* (Volosh.), *I. laticamerata* (Volosh.), and *I. kasiwazakiensis* (Husez. et Maruh.).

In the upper part of this subformation, i.e., in the *Buccella niigataensis–Islandiella helenae* Beds, the number of extinct species decreased to 12 - 15% (though these proportions are rather conventional).

In the upper subformation (*Retroelphidium hughesi–Epistominella pacifica* Beds) only three extinct species were found. They constitute 4 - 6% of the total assemblage, and among them only forms of *Retroelphidium hughesi* (Cush. et Grant) occur in abundance.

According to foraminiferal assemblages, the lower Ol'khovaya deposits accumulated in the sublittoral zone, similar to the upper Ol'khovaya Subformation. Bottom temperatures were rather stable with some fluctuations and low positive values.

On the whole, the microfauna from the Ol'khovaya Formation was transitional between the typical Pliocene and Quaternary assemblages, and its analogs are unknown on the Pacific coast of Russia.

DIATOMS

One hundred and fifty one samples were taken for the diatom analysis of the Ol'khovaya Formation deposits. Fossil diatoms were found in 48 samples from the Mutnaya and Ol'khovaya-1 River sections (Fig. 2). These samples yielded more than 100 species and subspecies. We distinguished three assemblages of different ages. The first assemblage is associated with the lower Ol'khovaya Subformation deposits exposed in the Mutnaya-B section. The second assemblage occurs in sediments of the same subformation exposed in the Mutnaya-A reference section. Finally, the third assemblage characterizes the upper Ol'khovaya Subformation deposits of the Ol'khovaya-1 River section (Figs. 2 and 3).

The Pliocene–Quaternary boundary is placed at the top of the Olduvai event of the Matuyama reversed polarity epoch and is dated at 1.66 Ma (Berggren *et al.*, 1985). The base of the Olduvai event (1.88 Ma) is correlated with the lower boundary of the diatom *Actinocyclus oculatus* Zone of the North Pacific, which is defined by the last occurrence of *Thalassiosira antiqua* (Koizumi, 1986; Koizumi and Tanimura, 1985; Akiba, 1986). Therefore, the Pliocene–Quaternary boundary corresponds to the very basal part of the *Actinocyclus oculatus* Zone. The lower boundary of the older *Neodenticula koizumii* Zone is defined by the disappearance of *Neodenticula kamtschatica*. Some species of *Thalassiosira*, characteristic of older zones, also became extinct near this boundary. Within the *Neodenticula koizumii* Zone, about 2 Ma, species *Pyxidicula zabelinae* disappeared, whereas *Pyxidicula horridus* and *Thalassiosira antiqua* became extinct at the top of the zone.

The lower boundary of the *Simonseniella curvirostris* Zone overlying the *Actinocyclus oculatus* Zone is

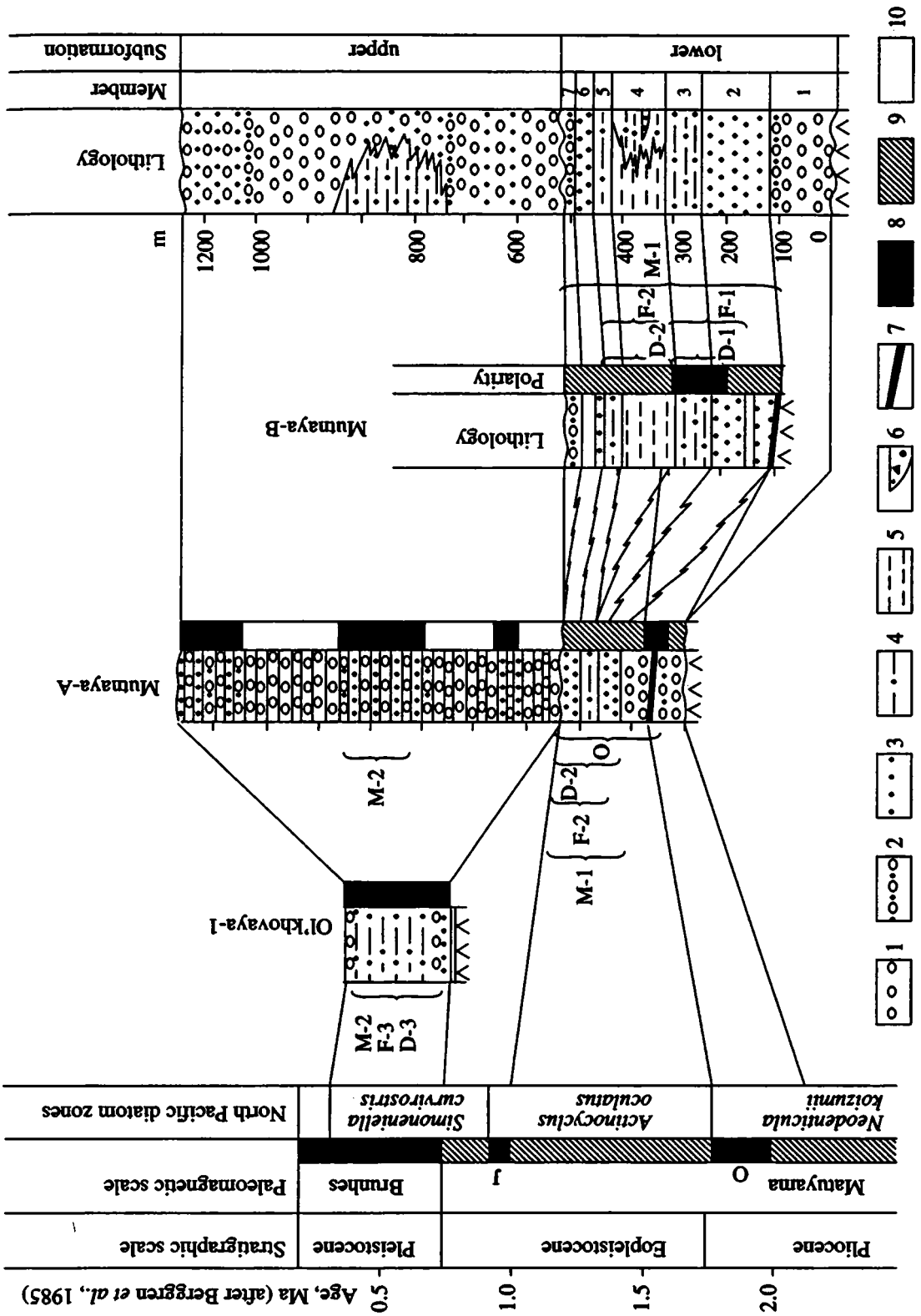

Fig. 2. Correlation of major sections of the Ol'khovaya Formation.
Fossil assemblages: mollusks – M-1 – *Quasisipho torquatus–Yoldia toporoki*; M-2 – *Nuculana minuta sachalinica–Chlamys beringianus*; foraminifers – F-1 – *Buccella conica–Islandiella excavata*; F-2 – *Buccella niigataensis–Islandiella helenae*; F-3 – *Retroelphidium hughesi–Epistominella pacifica*; diatoms – D-1, D-2, D-3.
O – deposits with ostracodes; 1 – conglomerates; 2 – interlayering of sandstones, conglomerates, and gravelstones; 3 – sandstones; 4 – sandy siltstones; 5 – siltstones; 6 – conglomerato-breccia of submarine slumps; 7 – tectonic contacts; 8 – normal; 9 – reversed; 10 – not defined.

Mollusks from the Ol'khovaya Formation

Sample number	Mollusk taxa	Distribution in the Ol'khovaya Formation	
		lower	upper
	Gastropoda		
001 BS	*Colisella cassis* (Eschscholtz)	+	
002 BA	*Margarites costalis* (Gould)	+	
003 BA	*M. helicina* (Phipps)	+	
004 BS	*M. vorticifera* (Dall)	+	
005 BA	*Solariella varicosa* (Mighels et Adams)	+	
006 BA	*Tachyrhynchus erosus* (Couthouy)	+	
007 BA	*T. reticulatus* (Mighels et Adams)	+	
008 BA	*Trichotropis bicarinatus* (Sowerby)	+	
009 BS	*T. coronatus* Gould	+	
010 BS	*T. insignis* Middendorff		+
011 BA	*Lunatia pallida* (Broderip et Sowerby)	+	
012 NB	*L. pila* (Pilsbry)	+	
013 BA	*Cryptonatica clausa* (Broderip et Sowerby)	+	+
014 +	*C. mammillata* (Deshayes)	+	
015 BS	*C. janthostoma* Deshayes	+	
016 +	*Volutopsius planus* Petrov	+	
017 BA	*Sipho* cf. *esychus* (Dall)		+
018 BA	*S. errones* (Dall)	+	+
019 BS	*S. halli* (Dall)	+	
020 BA	*S. jordani* (Dall)	+	
021 +	*Quasisipho torquatus* Petrov	+	+
022 BA	*Plicifusus kroyeri* (Moller)	+	
023 VB	*Neptunea amianta* Dall	+	
024 BA	*N. communis* (Middendorff)	+	
025 +	*N. humilis* Petrov	+	
026 VB	*N.* cf. *pribiloffensis* (Dall)		+
027 +	*N. unicostulata* Ilina	+	
028 BS	*N. lyrata* (Gmelin)		+
029	*N.* sp.	+	
030 VB	*Buccinum herzensternii* Verkrusen	+	
031 BA	*B. plectrum* Stimpson	+	
032 BA	*B. polare* Gray	+	
033 +	*Ancistrolepis bicostatus* Petrov	+	
034 BS	*A. okhotensis* Dall	+	
035 VB	*Beringius aleuticus* Dall	+	
036 VB	*Boreotrophon pacificus* (Dall)	+	
037 BA	*Admente couthouyi* (Jay)	+	
038 BA	*A. viridula* (Wood)	+	

(Contd.)

Sample number	Mollusk taxa	Distribution in the Ol'khovaya Formation	
		lower	upper
039 BS	*Aforia circinata* (Dall)	+	
040 BS	*Antiplanes voyi* Gabb	+	
041	*Rectiplanes* sp.	+	
042	*Plicifusus* sp.	+	
043 BA	*Oenopota harpularia* (Couthouy)	+	
044 VB	*O. metschigmensis* (Krause)	+	
045 +	*O. olchovensis* Petrov	+	
046 BA	*O. schantarica* (Middendorff)	+	
047 BA	*O. turricula* (Montagu)	+	
048 BA	*Boreoscala groenlandica crebricostata* (Sara)	+	
049 BA	*Cylichna occulta* (Mighels)	+	
	Bivalvia		
050 BA	*Leionucula tenuis expansa* (Reeve)	+	
051 BA	*L. inflata inflata* (Hancock)	+	
052 A	*L. inflata romboides* Scarlato	+	
053 BA	*Nuculana pernula* (Muller)	+	
054 A	*N. lamellosa lamellosa* (Leche)	+	
055 VB	*N. fossa* (Baird)	+	
056 BS	*N. minuta angusticauda* Scarlato		+
057 B	*N. minuta sachalinica* Scarlato		+
058 BS	*N. sadoensis* (Yokoyama)	+	
059 NB	*Yoldia* (*Cnesterium*) *excavata* Dall	+	
060 NB	*Y.* (*Cn.*) *toporoki* Scarlato	+	
061 +	*Y.* (*Cn.*) *toporoki olchovica* Petrov	+	
062 BA	*Yoldiella lenticula* (Moller)	+	
063 BS	*Megayoldia* (*Megayoldia*) *thraciaeformis* (Storer)	+	
064 BS	*M.* (*Portlandella*) *lischkei* (Smith)	+	
065 +	*M.* (*P.*) *olchovica* Petrov	+	+
066 BS	*Mytilus edulis* Linne	+	+
067 +	*M. edulis declivis* Petrov	+	
068 BA	*Crenella decussata* (Montagu)		+
069 BA	*Mucculus discors* (Linne)	+	
070 BA	*M. niger* (Gray)	+	
071 VB	*Chlamys* (*Chlamys*) *beringianus* (Middendorff)		+
072 +	*Ch.* (*Ch.*) cf. *coatsi* MacNeil	+	+
073 +	*Ch.* (*Ch.*) cf. *pilicaensis* Kubota	+	
074 BA	*Lyonsia arenosa arenosa* (Moller, 1842)	+	
075 BA	*Pandora* (*Pandorella*) *glacialis* Leach	+	

(Contd).

Sample number	Mollusk taxa	Distribution in the Ol'khovaya Formation	
		lower	upper
076 BA	*Periploma fragilis* (Totten)	+	
077 +	*Astarte* (*Elliptica*) *kamtschatica* Petrov	+	
078 BA	*A.* (*Tridonta*) *borealis borealis* (Schumacher)	+	
079 A	*A.* (*T.*) *borealis placenta* Morch	+	
080 +	*A.* (*T.*) *olchovica* Petrov	+	
081 VB	*A.* (*T.*) *rollandi* Bernardii	+	+
082 A	*A.* (*Nicania*) *montagui fabula* (Reeve)	+	
083 A	*A.* (*N.*) *montagui striata* Leach	+	
084 BA	*A.* (*N.*) *montagui vernicosa* Dall	+	
085 A	*A.* (*N.*) *montagui warhami* (Hancock)	+	
086 BA	*Hiatella arctica* (Linne)	+	+
087 BA	*Cyrtodria kurriana* Dunker	+	
088 BA	*Panomya arctica* (Lamarck)	+	
089 BA	*Thyasira gouldi* (Philippi)		+
090	*T.* sp.	+	
091 BA	*Axinopsida orbiculata* (Sars)	+	
092 BS	*Musella kurilensis kurilensis* Scarlato et Ivanova	+	
093 BS	*M. kurilensis litoralis* Scarlato et Ivanova	+	
094 BS	*Cyclocardia crebricostata* (Krause)	+	
095 NB	*C. erimoensis* (Tiba)	+	+
096 BS	*C. ventricosa*		+
097	*C.* sp.	+	
098 BA	*Liocyma fluctuosa* (Gould)	+	+
099 +	*Caliptogena lachtakensis* Petrov	+	
100	*C.* sp.	+	
101 BA	*Ciliatocardium ciliatum* (Fabricius)	+	
102 +	*C. olchovensis* Petrov	+	
103 BA	*Serripes groenlandicus* (Bruuiere)	+	+
104 BS	*Spisula* (*Mactromeris*) *voyi* (Gabb)	+	+
105 BA	*Macoma calcarea* (Gmelin)	+	
106 BA	*M. crassula* (Deshayes)	+	
107 BA	*M. moesta* (Deshayes)	+	
108 VB	*M. obliqua* (Sowerby)	+	
109 BS	*Siliqua alta* (Broderip et Sowerby)	+	
110 BA	*Mya* (*Mya*) *truncata* Linne	+	+
111 +	*M.* (*M.*) *truncata ovata* Jensen	+	
112 +	*M.* (*M.*) *truncata olchovica* Petrov	+	
113 BS	*M.* (*M.*) *priapus Tilesius*	+	
114 VB	*M.* (*Arenomya*) *elegans* (Eichwald)	+	+

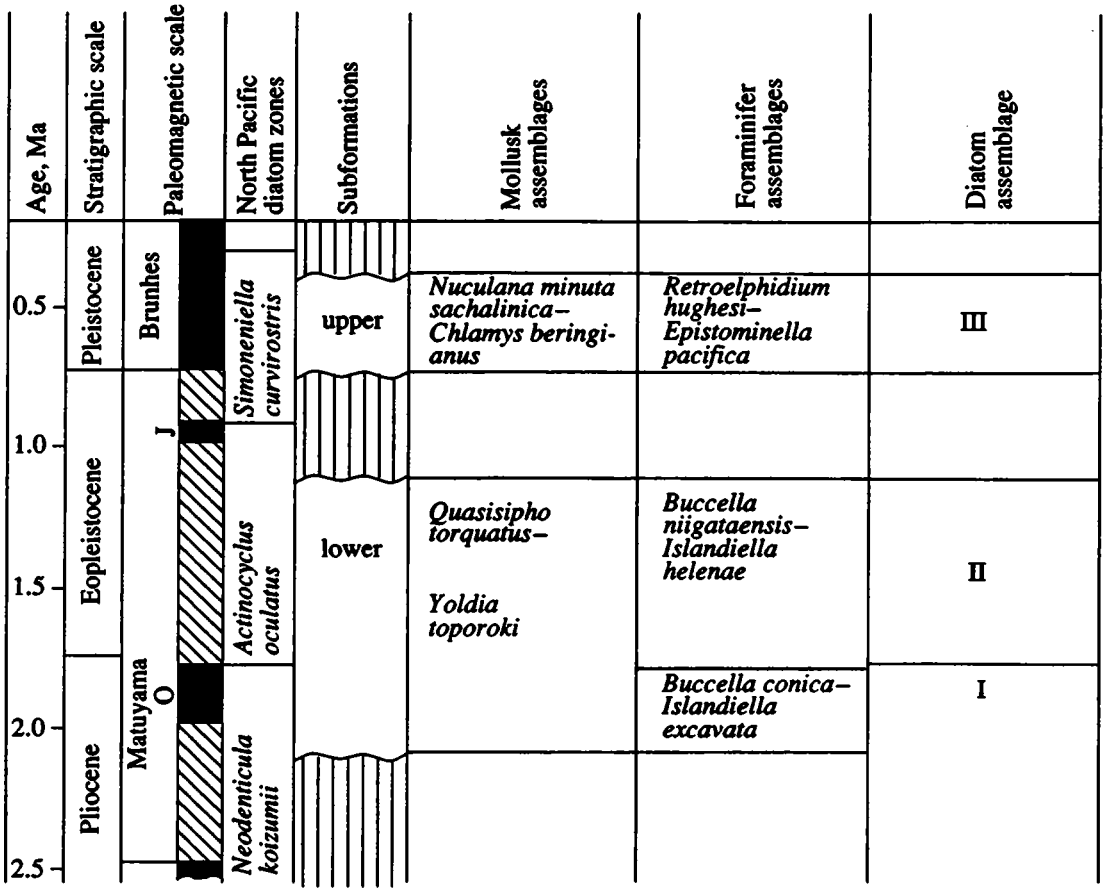

Fig. 3. Stratigraphic scheme of the Ol'khovaya Formation.

defined by the disappearance of *Actinocyclus oculatus*, which occurred at about 0.95 Ma in the subarctic region. Its upper boundary is marked by the extinction of *Simonseniella curvirostris* at 0.3 Ma (Koizumi and Tanimura, 1985; Koizumi, 1986).

The following diatoms passed from older deposits to the *Actinocyclus oculatus* Zone: *A. ochotensis*, *Simonseniella barboi*, *Thalassiosira gravida*, *T. gravida* f. *fossilis*, *T. oestrupii*, *T.* nut *nordenskioeldii*, *T. hidulus*, *Detonula confervacea*, *Bacterosira fragilis*, *Coscinodiscus marginatus*, *Porosira glacialis*, *P. punctata*, and *Neodenticula koizumii* (Koizumi and Tanimura, 1985; Oreshkina, 1989). These species, together with *Pyxidicula zabelinae* and *Thalassiosira antiqua*, are constituents of the first diatom assemblage from the lower Ol'khovaya Subformation (Mutnaya-B section). Therefore, this group may correspond to both the upper part of the *Neodenticula koizumii* and the lower part of the *Actinocyclus oculatus* zones. According to Petrov (1982), the deposits of the upper and lower parts of the section are reversely magnetized, whereas those of the middle part had a normal polarity (Fig. 2). Thus, we can suggest that the first diatom assemblage similar to that from the lower part of the *Actinocyclus oculatus* Zone corresponds to the Pliocene–Quaternary boundary in the upper part of the section, at the border of normally and reversely magnetized layers. Such a correlation is confirmed by the absence of *Simonseniella curvirostris* forms, which appeared in the middle part of the *Actinocyclus oculatus* Zone, above the top of the Olduvai event (Koizumi and Tanimura, 1985; Koizumi, 1986). At the same time, the occurrence of *Pyxidicula zabelinae* and *Thalassiosira antiqua* may be the result of their wider stratigraphic range in subarctic latitudes. For instance, according to Koizumi (1986), the last occurrence of *Thalassiosira antiqua* in different climatic zones of the North Pacific varied from 3.14 to 1.43 Ma. Moreover, we cannot exclude the possibility that these species were redeposited.

The second diatom assemblage is characterized by the disappearance of *Neodenticula koizumii* in its lower part and by the appearance of *Simonseniella curvirostris*, *Thalassiosira pacifica*, and species of *Asteromphalus* genus. Numerous and diverse representatives of *Thalassiosira*, as well as *Neodenticula seminae*, also occur there. The assemblage correlates to the lower Pleistocene part of the *Actinocyclus oculatus* Zone. According to Petrov (1982), the deposits yielding this assemblage are reversely magnetized, and therefore we can suggest that they accumulated during the upper part of the Matuyama epoch, above the Olduvai, but below the Jaramillo events.

The upper Ol'khovaya Subformation deposits, which contain the third diatom assemblage, have a normal polarity (Petrov, 1982). Therefore, the diatom assemblage lacking *Pyxidicula zabelinae* and *Thalassiosira antiqua*, but containing *Neodenticula seminae*, and *Simonseniella curvirostris*, which became extinct at 0.3 - 0.35 Ma, as well as *Actinocyclus ochotensis* and rare *A. oculatus* evidently corresponds to the uppermost part of the *A. oculatus* Zone associated with the Jaramillo normal polarity event. The boundary between the *Simonseniella curvirostris* and *Actinocyclus oculatus* zones is defined by the last occurrence of the *Actinocyclus oculatus* and coincides with the upper boundary of the Jaramillo event (0.91 Ma). But, according to some records (Koizumi and Tanimura, 1985; Sancetta and Silvestri, 1986), rare specimens of *Actinocyclus oculatus* occur above this level in a number of DSDP sites and piston cores of the North Pacific. Therefore, we can also suggest that the sediments corresponding to the Jaramillo event and the reversely magnetized interval in the uppermost part of the Matuyama epoch are absent in the Ol'khovaya Formation. Thus we cannot exclude the possibility that its third diatom assemblage may correlate to the *Simonseniella curvirostris* Zone at the base of the Brunhes paleomagnetic epoch (Fig. 2).

On the whole, the diatom flora from the Ol'khovaya Formation consists mainly of planktonic neritic and oceanic species, sublittoral plankton, and benthos. The neritic plankton prevails, whereas oceanic species are more scarce. The number of sublittoral and benthic diatoms is also low, though at some levels they are rather numerous. The majority of Ol'khovaya diatoms are referred to boreal and arctic-boreal Pacific species. The flora was formed in a highly productive, relatively shallow shelf zone connected with the open sea and characterized by cold waters covered by ice in the winter. The sea level fluctuated, as is indicated by the heterogenous composition of the assemblages (the second one has a relatively deep-water origin) and by shallow-water facies of sediments barren of diatoms (the lower beds and upper part of the formation). Judging from the presence of relatively thermophilic forms, i.e., *Thalassiosira lineata*, *T. leptopus*, and others in the second assemblage, a slight warming occurred between the Olduvai and Jaramillo events.

CONCLUSION

The study of the Ol'khovaya Formation in eastern Kamchatka revealed certain changes in mollusk, foraminifer, and diatom assemblages, which served as the basis for distinguishing pertinent zones and faunal beds (Fig. 3). Moreover, diatom zones were correlated with the zonal scale of the boreal region. At the same time, paleomagnetic orthozones in the section were referred to units of the magnetochronologic scale. This enabled us to formulate a fairly reliable definition of the Neogene–Quaternary boundary (1.65 Ma) in the lower part of the formation and of the Eopleistocene–Pleistocene boundary (0.7 Ma) at the base of the upper subformation. The changes in foraminifer and diatom assemblages are confined precisely to these boundaries, and the most pronounced faunal changes occurred near or at the Pliocene–Eopleistocene boundary (1.65 - 2 Ma). The study of benthic assemblages from the transitional Pliocene–Quaternary beds in northeastern Asia demonstrated that the boreal faunal groups, which were formed at that time, did not significantly change until recently, a fact which shows their high stability under stress situations, i.e., climatic fluctuations, and others.

Within the Ol'khovaya sediments dated as the late Pliocene–Pleistocene (2.5 - 0.5 Ma), two sedimentary cycles separated by a hiatus were distinguished. According to diatom and paleomagnetic data, the hiatus was not shorter than the upper part of the Matuyama epoch, which corresponds to the lower part of the diatom *Simonseniella curvirostris* Zone, and lasted about 300 ka. Deposits of the formation yield ice-rafted pebble and traces of volcanic eruptions.

A comparison of fossil assemblages from the Ol'khovaya Formation with the older Pliocene fauna and flora of Karaginskii Island revealed a certain climatic cooling trend in the North Pacific during the late Pliocene and Quaternary. Some warming, evidently, occurred in the interval of 1 - 1.5 Ma. It should be noted that Ol'khovaya deposits deformed strongly into faulted folds. These deformations took place in the Middle Pleistocene, when, evidently, the recent Kamchatka structures were formed.

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Silurian Chitinozoa from Western Guinea

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In recent decades, western Guinea was an object of geological mapping and mineral prospecting. Studies of geology are, thus, of major practical and theoretical interest. However, the age, stratigraphic subdivision, and correlation of sedimentary sequences of the region are still under debate.

A number of geologists first studied the geology of the region. Hubert (1919) compiled a geological map on a 1 : 500000 scale. Sinclair (1928) was the first to discover Silurian graptolits in black, pyrite-containing claystones 10 km to the west of Telimele. Chetelat (1939), while conducting field studies, discovered Silurian and Devonian fossils, which were used for the stratigraphic subdivision of the Lower Paleozoic. Arnould and Guilaume (1959) described all the Silurian deposits in northern Guinea in the Telimele Shales Formation.

The Telimele Formation deposits occur chiefly in the central part of the Paleozoic sedimentary basin. They form a platform cover of the west Guinea synеclise in the west of the republic. The deposits conformably overlie the Ordovician Pita Formation sandstones and outcrop along river valleys. The Telimele Formation is composed of claystones with subordinate beds and sandstone lenses.

Claystones are dark-grey, almost black, more rarely grey. They are commonly micaceous and loosely cemented. The rocks mainly consist of sericite, sericite–chlorite, or hematite–chlorite fine-grained aggregates. Clastic material is represented by pelitic and fine-grained quartz. Rocks are thin-platy and display thin parallel bedding. Sulphides occur on bedding planes.

The siltstones are fine-grained, clayey, dark grey and light grey, and compact. The bulk of the rock consists of quartz grains. The cement is sericite and chlorite, sometimes clayey.

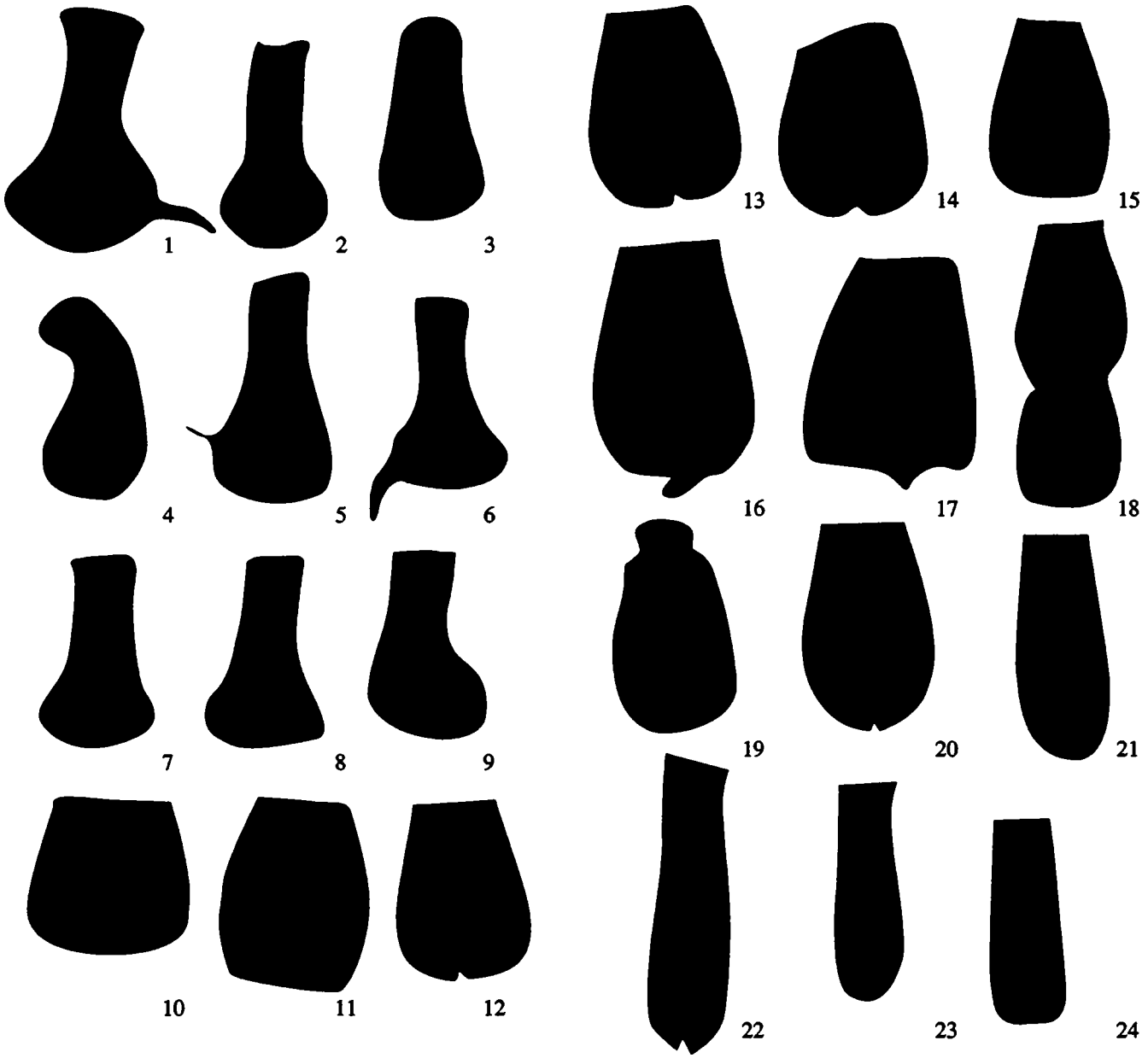
Sandstones are dark grey, variably-grained, and massive. Poorly rounded grains of angular or irregular form predominate. The cement is basal, sericite by composition, containing a dissemination of fine-grained hematite.

The total thickness of the Telimele Formation deposits gradually increases to the northwest, reaching 550 m.

The Telimele Formation sequences show noticeable lateral and vertical compositional variations due to varying amounts of siltstones or fine-grained sandstones. For example, in the southwest, the formation is compositionally uniform or has a two-unit structure. In the latter case, the lower unit, 100 m thick, consists of claystones with interbeds of siltstones and the upper one, 300 - 350 m thick, is composed of claystones. In the west, the formation has a three-unit structure. Here, a unit of claystones and siltstones 180 m thick is overlain by a 40 - 60-m-thick sequence of claystones, siltstones, and sandstones. Higher, a sequence of siltstones and claystones with isolated beds of quartz sandstones occurs. In the northeastern part of the region, the Telimele Formation deposits become more coarse-grained and consist of siltstones with interbeds of sandstones.

The brief lithologic description of rocks of the Telimele Formation given above indicates sedimentation of the Silurian deposits in a temporarily closed basin. The basin was characterized by stabile environments, low hydrodynamics, and fairly diverse organic life. Decomposition of organic remains resulted in the reduction of conditions favoring pyrite formation at the bottom of the basin, which caused the black color of deposits. Beds of quartz conglomerates and lenses of siltstones indicate more turbulent conditions of sedimentation in the middle and at the end of the Silurian. The southeastern part of the region was the shallowest area of the basin, where a decrease in thickness and coarser sediments occur.

The Silurian age of the Telimele Formation is inferred from numerous finds of graptolites, brachiopods, and trilobites, as well as from its stratigraphic position above the Ordovician Pita Formation and below the Devonian Faro Formation. Graptolite and brachiopod assemblages suggest a plausible presence of Llandoverian and Ludlowian, and presumable Wenlockian deposits within the Telimele Formation sequence. To obtain more complete paleontological characteristics and, thus, an additional age substantiation of the Telimele Formation deposits outcropping in the vicinity of the Bake town and the villages of Yumbuya and Tombuya, where they are composed of black-



1, 2 – *Agathochitina primitiva* (Eisenack) Tsegelnjuk, 3 - 5 – *Agathochitina dilatata* Tsegelnjuk, 6, 7 – *Ancyrochitina deltoidea* Tsegelnjuk, 8 – *Fungochitina incompta* Tsegelnjuk, 9 – *Discochitina depressa* Tsegelnjuk, 10 – *Bursachitina cylindrica* (Taugourdeau et Jekhowsky) Tsegelnjuk, 11 - 19 – *Linochitina cingulata* (Eisenack) Tsegelnjuk, 20 – *Clavachitina conica* (Taugourdeau et Jekhowsky) Tsegelnjuk, 21 - 23 – *Clavachitina longiuscula* Tsegelnjuk, 24 – *Conochitina pauca* Tsegelnjuk.

grey, almost dark micaceous, massive claystones, a microfaunal analysis was performed.

The following forms of Chitinozoa were determined in these rocks: *Linochitina cingulata* (Eisenack) Tsegelnjuk, *L. deminuta* Tsegelnjuk, *Clavachitina longiuscula* Tsegelnjuk, *C. conica* (Taugourdeau et Jekhowsky) Tsegelnjuk, *Conochitina pauca* Tsegelnjuk, *Bursachitina cylindrica* (Taugourdaeu) Tsegelnjuk, *Agathochitina primitiva* (Eisenack) Tsegelnjuk, *Ancyrochitina deltoidea* Tsegelnjuk, and *Fungochitina incompta* Tsegelnjuk.

Abundant and well-preserved specimens of more than 10 Chitinozoa genera from Silurian claystones of the Guinea Republic have been recovered for the first time.

Brief descriptions of some Chitinozoa species are given below.[1]

Agathochitina primitiva (Eisenack, 1964) Tsegelnjuk, 1982

The plate, Figs. 1 and 2. Black tests up to 180 μm long. The chamber is almost spherical, ranging in size

[1] Stored at the Institute of Geological Sciences, Ukrainian Academy of Sciences, Department of Fossil Flora.

from 73 μm to 98 μm. The width of the neck is as much as 36 μm. It is somewhat wider towards aperture, reaching 50 μm. The flexure is sharp. The walls and base of the chamber are convex. The basal margin, with processes up to 40 μm long, is rounded.

Distribution. The Lower Silurian, the Wenlockian Stage; the Restevo, Demshin, and Vrublevtsy subformations of Podoliya; and Slite Beds of the Gotland Island.

Agathochitina dilatata Tsegelnjuk, 1982

The plate, Figs. 3 - 5. Black tests with a rounded, somewhat elongated chamber. The length of tests ranges from 125 to 156 μm, the width of the apertural part of the neck is 35 - 42 μm, and the width of the neck's base is up to 58 μm. The largest diameter of the chamber is confined to its middle part and is equal 92 μm. The flexure is gradual. The basal margin of the chamber commonly bears isolated exoderm processes up to 5 - 8 μm long.

Distribution. The Lower Silurian, the Wenlockian Stage; and the Demshin, and Vrublevtsy subformations of Podoliya.

Ancyrochitina deltoidea Tsegelnjuk, 1982

The plate, Figs. 6 and 7. Black tests with a narrow (18 - 25 μm) neck and a large (78 - 106 μm) conical chamber. The length of test varies from 130 to 142 μm. The surface of the chamber bears isolated spines near its base. They are 10 - 17 μm long and 5.5 μm wide. The test wall is smooth, and the flexure is rather distinct. The chamber base is slightly convex, and the basal margin is rounded.

Distribution. The Lower Silurian, the Wenlockian Stage; the Ternava Formation and the lower part of the Muksha Subformation of Podoliya.

Fungochitina incompta Tsegelnjuk, 1982

The plate, Fig. 8. Black elongated (120 - 160 μm) tests with a wide (70 - 84 μm) fungiform chamber (of 130 - 142 μm long) and a narrow neck (22 - 32 μm). The flexure is sharp. The chamber walls are smooth. The base is flattened, and the basal margin is rounded.

Distribution. The Lower Silurian, The Wenlockian Stage; and the Demshin and the lower part of Vrublevtsy Subformation of Podoliya.

Discochitina depressa Tsegelnjuk, 1982

The plate, Fig. 9. Comparatively small tests (115 - 120 μm) of black color. The chamber is deformed. Its size varies from 68 to 92 μm. The cylindrical neck is up to 33 μm long. The flexure is smooth. The basal margin of the chamber is slightly rounded, and the chamber walls are smooth. The basal margin often bears narrow, spine-like processes 15 - 20 μm long, which are wide at the base and gradually taper distally. Their points are often rounded.

Distribution. The Lower Silurian, the Wenlockian Stage; upper part of the Furmanovka and Ternava formations of Podoliya.

Bursachitina cylindrica (Taugourdeau et Jekhowsky, 1960) Tsegelnjuk, 1982

The plate, Fig. 10. Short (70 - 92 μm) black tests 80 - 100 μm wide. The width of the aperture is 57 - 82 μm. The walls of the tests are straight. The base of the tests is slightly convex, and the basal margin is rounded.

Distribution. The Lower Silurian, the Llandoverian Stage; the lower part of the Restevo Subformation of Podoliya; the fourth and fifth Silurian Saharan zones.

Linochitina cingulata (Eisenack, 1937) Tsegelnjuk, 1982

The plate, Figs. 11 - 19. Black tests 136 - 180 μm long are slightly widened in the lower part (80 - 130 μm) and narrow near the aperture (45 - 75 μm). The base is almost flat, the basal margin is rounded, and the walls are smooth. The surface sometimes bears tiny tubercles or spines.

Distribution. The Lower Silurian, the Wenlockian Stage; the Vrublevtsy and Sursha subformations of Podoliya; Slite and Mulde beds of Gotland Island.

Clavachitina conica (Taugourdeau et Jekhowsky, 1964) Tsegelnjuk, 1982

The plate, Fig. 20. Black forms with a length of 235 - 385 μm and a width of 110 - 173 μm, inflated and quickly tapering towards the aperture of neck (43 - 110 μm). The walls and base are convex, and the basal margin is rounded.

Distribution. The Lower Silurian, the Wenlockian Stage; the Sursha and Vrublevtsy subformations of Podoliya; upper part of the Visbi Beds of the Gotland Island.

Clavachitina longiuscula Tsegelnjuk, 1982

The plate, Figs. 21 - 23. Black cylindric tests are 268 - 390 μm long and 65 - 80 μm wide, narrowing towards the neck aperture to 55 - 60 μm. The walls are straight, the base is convex, and the basal margin is rounded.

Distribution. The Lower Silurian, the Wenlockian Stage; the Sursha, and the lower part of Muksha Subformation of Podoliya.

Conochitina pauca Tsegelnjuk, 1982

The plate, Fig. 24. Relatively small black cylindric tests with a length of 80 - 170 μm, a width of 55 - 62 μm, and an aperture width of 25 - 48 μm. The walls are straight, the base is nearly flat, and the basal margin is rounded.

Distribution. The Lower Silurian, the Wenlockian Stage; the upper part of the Sursha Subformation of Podoliya; the upper part of the Visbi Beds of Gotland Island.

Chitinozoa from Silurian deposits of the western Guinea Republic are identical to forms from the Lower Silurian (Wenlockian) deposits of Podoliya, the Ukraine (Tsegelnjuk, 1982), and the Siberian platform (Obut, 1973). It should be noted, however, that the Guinean Chitinozoa are more variable in main morphological features than the forms from the East European and Siberian platforms, which apparently reflects specific environments. Further studies of the Lower Silurian Chitinozoa of Guinea will provide a basis for elaborating a detailed stratigraphic scale for this region and for correlating distant sections.

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Seismostratigraphic Criteria for Distinguishing Marine from Continental Facies

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Abstract – Three groups of seismostratigraphic criteria allow us to reliably discriminate between marine and continental facies. The first group consists of paleogeomorphological elements, the second, accumulative bodies, and the third, patterns of seismic records. Their integrated application provides the most adequate solution to the problem.

The correlation of continental deposits with coastal and marine sediments is frequently problematic, which is the reason why many stratigraphic and paleontological papers have been devoted to this topic. Seismostratigraphic data has made a significant contribution in recent years, permitting us to propose a number of important criteria for correlation, resulting, together with biostratigraphic and paleomagnetic information, in more reliable correlational schemes.

There are three groups of seismostratigraphic criteria that have to be used in the correlation of marine and continental facies. The first includes paleogeomorphological elements that are observed in time seismic reflection sections. The second group is based on the form of the accumulative bodies, allowing us to recognize their genesis. The third group comprises seismic reflection patterns. The combined use of these three groups of criteria is key to resolving the problem of the discrimination and correlation of different facies widespread in epicontinental basins. The latter are divided into two types. The first are active basins with a synsedimentary subsidence. They are distinguished by an accumulation of marine or continental facies in accordance with the depth of the basin. The second type is represented by passive basins, where the sedimentation is only affected by sea level fluctuations. Sedimentation in these basins occurs only on the shelves and is restricted by the shoreline.

1. PALEOGEOMORPHOLOGICAL CRITERIA

Paleoslopes can be most reliably recognized in sedimentary sequences. They are outlined by the marine onlap discordance between the paleoslope surface and the overlying layers. The paleoslopes under the influence of terrigenous sedimentary deposition are buried under the progradational bodies (clinoforms) and reversed clinoforms. Disconformities also occur at their base. When the slopes' inclination exceeds 5°, their surfaces are complicated by sheet erosion, which truncates the layers of slope deposits. Paleoslopes reliably indicate a relatively deep marine or lacustrine sediment genesis.

Land paleoincisions are reliable paleogeomorphological indicators of terrestrial conditions. They are distinguished by the erosion of the underlying deposits and burial under the overlying deposits, whose layers display an onlap fill against the slopes of paleoincisions. The basin floor was situated above the sea-level during the entrenchment, and sedimentation was, in fact, absent. Land paleoincisions should be distinguished from submarine canyons. The former are characterized by maximum relative depths in their middle and upper parts. Land paleoincisions become less evident near their mouths and completely degrade in the adjoining sea floor areas. Submarine paleocanyons are usually related to paleoslopes and fade out on the paleoabyssal plain. Deep-sea paleochannels created by deep-water currents are linear, allowing us to discriminate them from complicated branching systems of land paleoincisions.

Truncated (discordant) paleosurfaces are principally related to upward vertical or lateral (compressional) tectonic movements and marine transgressions widespread in the sedimentary cover. Flat, truncated surfaces are produced by regular wave action on the shelf. On land, the truncated surfaces are complicated by smaller incisions due to irregular denudation and, chiefly, erosional activity. Thus, the form of the truncated surfaces provides impartial paleogeomorphological criteria. Flat, discordant surfaces reflect shelf conditions at a definite stratigraphic level, and uneven, incised surfaces are peculiar to continental conditions.

Basal (coastal) onlap arises near the coastline of a basin and is directed landward. This feature is related to the sea-level fluctuations. The coastal onlap gradually shifts landward during a transgression and seaward during a regression. A coastal onlap can be distinguished from the synsedimentary tectonic onlap, and the onlap fill of a paleorelief burial. A synsedimentary

tectonic onlap is characterized by a divergence of seismic reflectors towards a basin, indicating layer thickening, while their facies composition is constant or changes slightly. On the contrary, a coastal onlap displays sedimentary layers strictly parallel to each other, and their facies composition changes sharply from coastal marine to inner shelf deposits. A coastal onlap differs from onlap fill in its much larger magnitude (a few hundred meters to a few kilometers versus a few dozen to hundreds of meters) and in the grade of the underlying sedimentation surface (a few degrees versus dozens of minutes). A coastal onlap is a reliable seismostratigraphic criterion for distinguishing the coastal-marine facies and positioning the paleocoastlines. Unfortunately, this is rarely found in actively subsiding tectonic basins, but is a typical feature of passive basin parts where the sedimentation is dependent on sea-level fluctuations.

Onlap fill is widespread in sedimentary sequences. It appears in relatively deep troughs beyond the shelf break, where wave processes have no influence. Sedimentary material is transported there by gravity as a viscous liquid. Trough slopes and bottom topography impede currents, which results in the onlap fill of the underlying relief. Onlap fill may occur in land and submarine erosional cuts. This is a reliable criterion for paleorelief reconstruction. The height, magnitude, and shape of the paleorelief are important parameters distinguishing the origin of the sediments.

2. ACCUMULATIVE (SEDIMENTARY) BODIES

Two genetic types of sedimentary bodies are spread over the deepest parts of basins. The first is represented by lateral fill sequence bodies. They originate from the input of terrigenous sediment, when the material is distributed by gravity as a viscous liquid. The strata of the bodies of the lateral fill sequence are characterized by stable thicknesses, and terminate with an onlap or with a wedgelike thinning toward the paleoslope break (reversed clinoform). The second genetic type is represented by hydrogenic sequences ("particle-by-particle" sedimentation). They are formed by sediments precipitating from the water layer, located either in inner ocean parts or in basins where clastic terrigenous input ceased for some periods due to specific sedimentation conditions. The hydrogenic sequences of inner oceanic regions are characterized by moderate (usually up to several hundreds of meters) thicknesses that remain constant over a wide area. They blanket the surface of the oceanic basement. However, in some depressed areas of the basement, the sediments of a hydrogenic sequence may be thicker (many hundreds of meters) due to gravity flows analogous to lateral fill sequences. The enlarged thickness of a hydrogenic sequence (up to a few hundreds of meters) can be observed on submarine uplifts consisting of a carbonate ooze that does not accumulate in the deeper parts of the neighboring abyssal plains. The bottom oceanic currents strongly influence the distribution of the hydrogenic sequences, leading to a differentiation in their thickness. The bodies of depressional sequences in deep basins are usually moderate in thickness (normally dozens of meters, and, more rarely, several hundreds of meters). Their strata grade with a sharp thickening into shelf carbonate, terrigenous, and terrigenous-carbonate prograding deposits (clinoforms) of paleoslope areas.

Sedimentary sequences of deep basins on the slope areas of high sedimentary influx sharply increase in thickness toward the abyssal plain. They form a sedimentary wedge, whose top coincides with the shelf break. This can be considered a reversed clinoform, because the strata thin out toward the basin margin. The reversed clinoform represent the peripheral part of a lateral deepwater fill sequence in the zone of its junction with the shelf sedimentary facies.

Strata of the carbonate shelf become thinner beyond the shelf break, thus forming a normal clinoform. It is also connected with strata of the basinal floor sequence. The carbonate shelf is often complicated by a barrier reef in the transitional zone to normal clinoform. Strata in a normal clinoform thin from many hundreds of meters near the paleoshelf break to a few dozens of meters near the paleocontinental rise.

When sediments become quick, lateral sedimentation occurs, leading to an accumulative growth of continental slope. Areal terrigenous clast transportation results in progradational bodies (clinoforms) that quickly pinch out downdip (fondoform) and updip (undaform). They can be classified into two types. The first of them is represented by sedimentary wedge covers (transgressive clinoforms), having somewhat thinner shelf counterparts. The second type of progradational bodies is represented by cyclic sedimentary wedges, which show that toplap patterns updip eventually pinch out on the shelf. The thickness of the progradational bodies, which are a few kilometers across, normally does not exceed a few hundreds of meters. Through the fondoform below, they grade into thin beds of depressional sequences.

Linear terrigenous clastics transportation results in submarine fans. They build up a continental slope and bury the neighboring abyssal areas. Submarine fans change along and across their trends into fondoforms, and then into thin beds of a depressional sequence. The thickness of submarine fans deposits varies between several hundreds of meters to a few kilometers. They sometimes overlap each other, forming a continuous belt parallel to the continental slope; across terrigenous scarps, they transform into thin beds of depressional sequences.

Shelfs occupy most of the epicontinental basins. They are balanced between sedimentation and submarine erosion. The balance disturbed by tectonics or sea-level fluctuations may result in the onset of a sedimentation episode. When the terrigenous input or biogenic activities are high, sedimentation itself destroys the balance and forms local accumulative bodies complicating the general pattern.

Elementary deltas develop mainly in the inner shelf under the influence of a high input of terrigenous clastics supplied by rivers. In the case of quick sediments, the sedimentation progrades in laterallike direction. The resulting cross-bedded bodies resemble those of slope progradation but are significantly thinner (up to a few tens of meters). They grade into thinner, synchronous prodeltaic beds and shelf deposits (centimeters to a few meters).

Conditions for short-term accumulations of bars arise due to wave-cut activity in local coastal areas of the shelf zone. The thickness of such bodies does not exceed a few dozens of meters, and the length is as much as several kilometers. Their common shape is isometrical, with convex top and flat basal surfaces. The shelf sediments that accumulate simultaneously fall, in general, below the resolution capacity of conventional seismics.

Reefs are complicating elements of the carbonate shelf platform. Barrier reefs are located in the zone bordering the shelf break and are the largest reefs. Reefs have a significant thickness, a slightly convex top, and flat basal surfaces. Anticlinal folds are often observed in the shelf sediments located above a reef. The poor quality of the seismic record is typical of reef structures. Inner reflectors are, in this case, distorted and discontinuous.

Elementary deltas, bars, and reefs are accumulative bodies that indicate coastal and shelf sedimentation paleoconditions. They provide reliable information on the genesis of the enclosing sedimentary cover.

Subaerial sedimentation in epicontinental basins occurs solely as a result of tectonic subsidence. Accumulative bodies are complementary elements, serving as indicators of sedimentary rock genesis.

River channel bodies are notable for their significant thickness (up to several dozens of meters) and extended sinuous shape. Their top surface is slightly convex, while the basal one is flat or slightly concave. The inner reflectors are discontinuous and distorted.

Subaerial fans have a notable thickness (dozens to a few hundreds meters) and are elongated. They have a convex top surface and a flat bottom. These sedimentary bodies are seismically transparent or show discrete chaotic reflections. Unlike their submarine counterparts, subaerial fans are unrelated to continental slopes, usually located in the peripheral areas of the basin floor.

3. SEISMIC PATTERN ANALYSIS

Patterns of seismic records are very useful in determining the facies composition and structure of sedimentary covers. However, this tool should be used carefully, because seismic patterns are strongly influenced by many sedimentological factors and subsequent diagenetic processes. Therefore, similar seismic patterns in different regions may reflect different sedimentation conditions and sediment compositions, but, to the contrary, different seismic patterns may correspond to similar facies and depositional environments. Therefore, seismic records should be considered together with the above listed paleogeomorphological criteria, pertinent to accumulative bodies, as well as with drilling results. Specific regional and local criteria should be understood and used only in limited areas. Some characteristics of seismic records on deposits of different genesis are presented below.

Deep-water trough deposits are characterized by two seismic records, namely by zones of strong reflectors of high amplitude in combination with transparent seismic units (Fig. 1). The vertical transition between them may be rather sharp. However, sometimes there is a gradual transition through semitransparent seismic horizons. The boundary between transparent and stratified seismic horizons lacks age and lithological significance. It is stratigraphically diverse and does not indicate contact relations between sedimentary units. Transparent seismic horizons are often encountered in the hydrogenic sequences of inner oceanic regions. Their formation may possibly be predetermined by a specific type of sedimentation. During the gradual deposition of sedimentary cover beds, sharp acoustic impedance contrasts are impossible, and the medium becomes unfavorable for the detection of seismic reflections in it, thus forming transparent seismic patterns. On the contrary, when sedimentation pulses, or the deposition is frequently interrupted by nondepositional periods, sharp acoustic impedance gradients result, which stimulate the appearance of high amplitude, continuous reflectors. The depressional sequence is usually marked by a strong reflector of high amplitude, which serves as the reference seismic horizon (Fig. 2).

Paleoslope sediments are distinguished by discontinuous weak reflections (Fig. 3). Oblique or sigmoidal reflector configurations are typical in this case. The strongest reflectors are confined to the tops and bottoms of wedgelike progradational bodies, and extend into neighboring paleoslope areas. In paleobasins, they merge with strong reflections of depressional sequences. Wedge cyclic units are notable for their weak reflectors, associated with rare, strongly reflecting horizons. The reversed clinoform deposits, especially on very steep slopes, are characterized by rare and chaotically oriented reflections. In general, such seismic patterns resemble the structure of a "rubbish heap".

Paleoshelf deposits are distinguished by groups of strong, continuous reflectors among horizons with weak reflections. Normally, the first groups are related to low energy deposits, usually represented by clayey sediment. Thicker units with weak reflections correspond to coarse terrigenous clastic rocks or biogenic, high energy sediments of a carbonate paleoshelf.

Seismic records of *paleoterrestrial deposits* are usually less clear (Fig. 4). Continuous reflectors occur occasionally in them, normally typical only of lacustrine deposits. It is difficult to trace synphase axes of reflections in such records because they degrade into short, uncorrelatable reflectors. The latter often have chaotic inclinations on seismic records of molasses

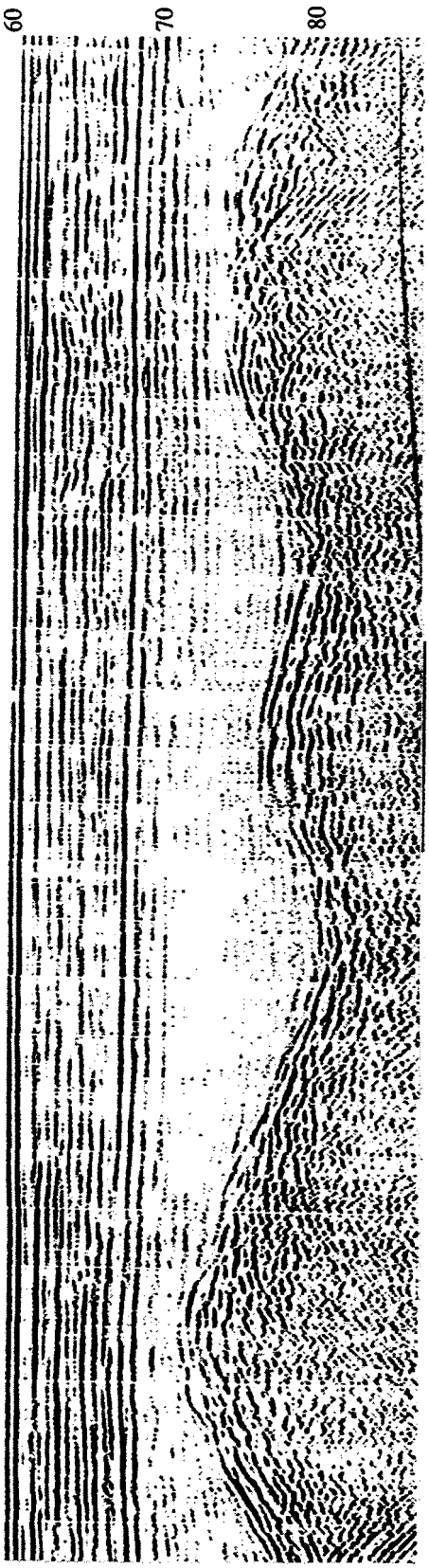

Fig. 1. A fragment of seismic record from the Sierra Leone Atlantic margin showing transparent and stratified seismic units of paleobasin deposits.

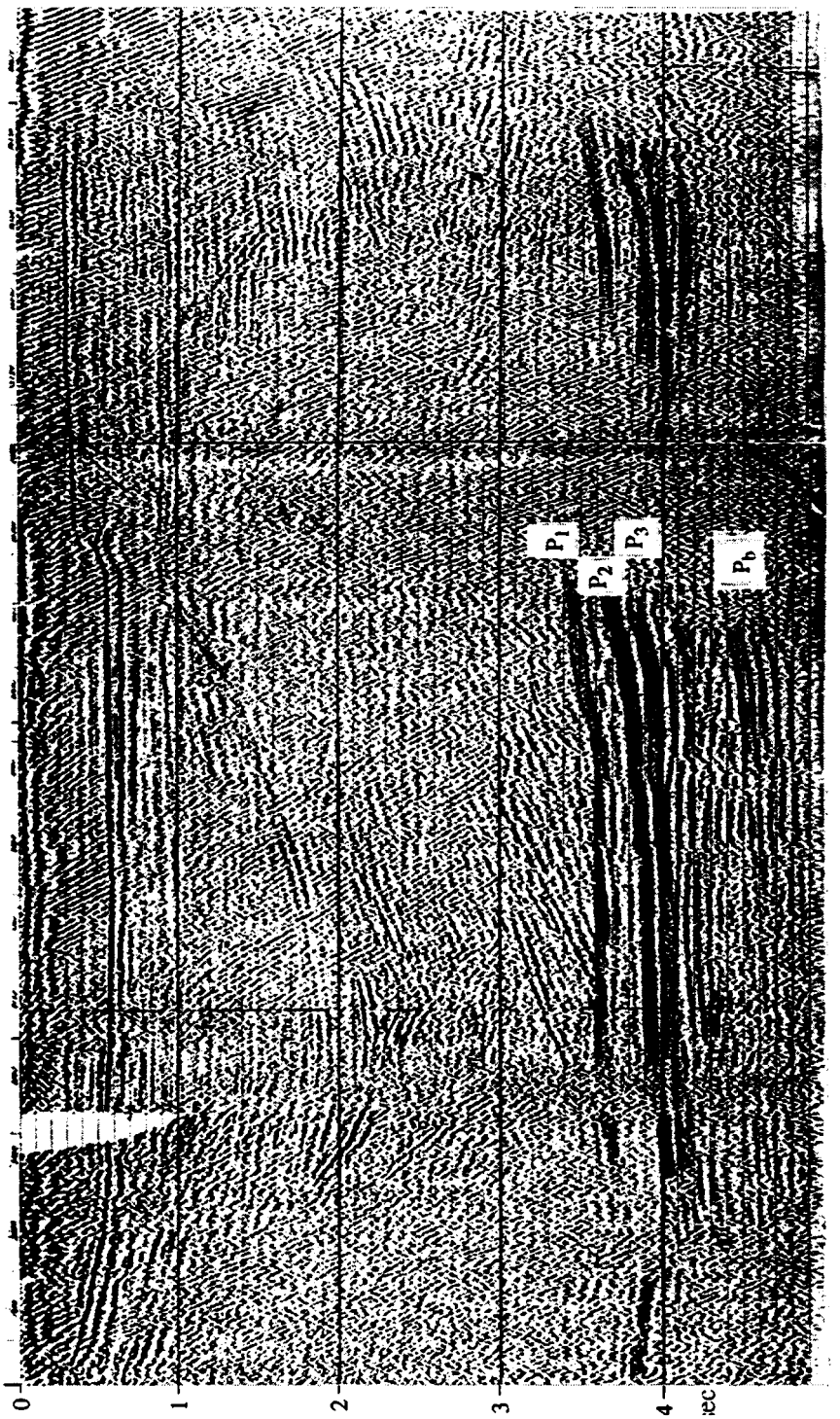

Fig. 2. A fragment of seismic record from the inner part of the North Caspian onshore basin showing strong reflectors related to rocks of depressional sequence. P_1, P_2, and P_3 are beds below salt deposits of the Paleozoic sequence; P_b – basement.

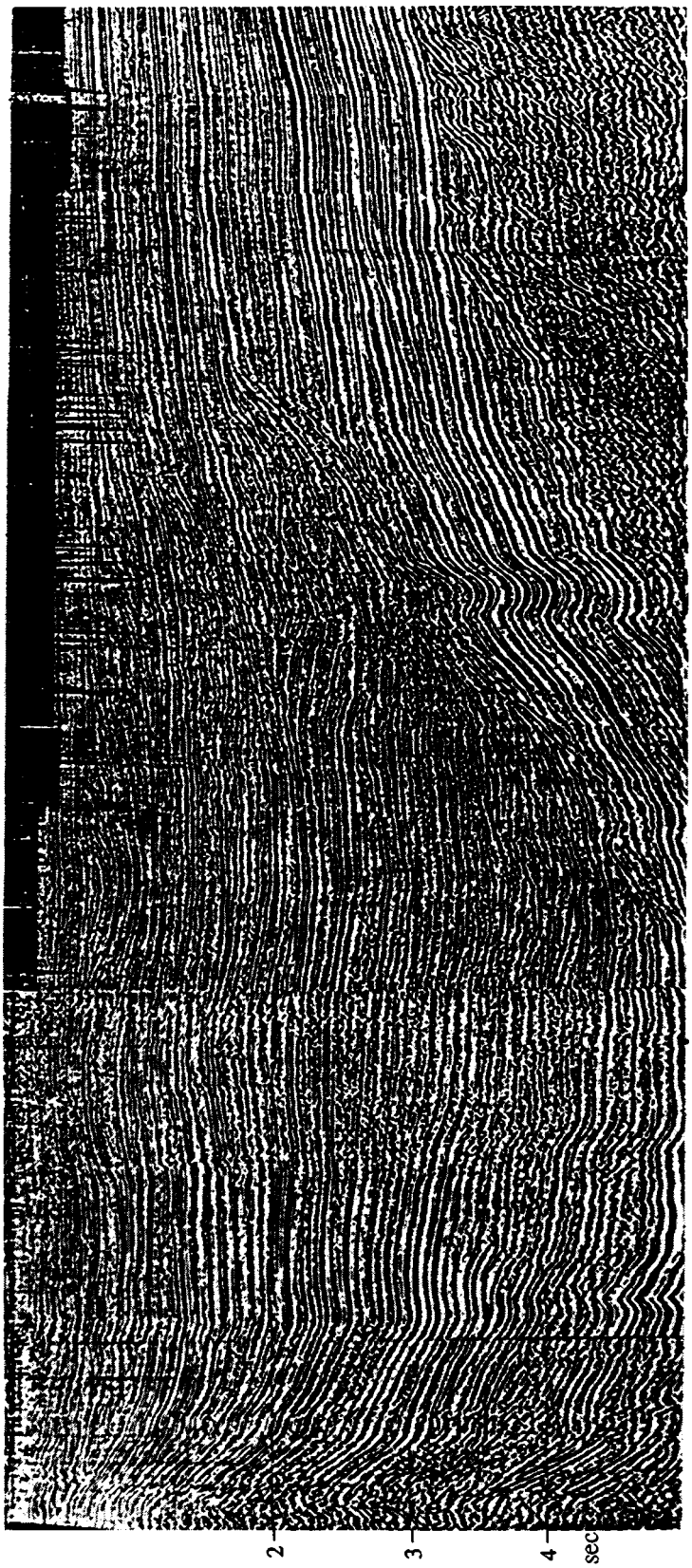

Fig. 3. A fragment of seismic section from the western portion of the Scythian Platform demonstrating patterns in the paleoslope deposits.

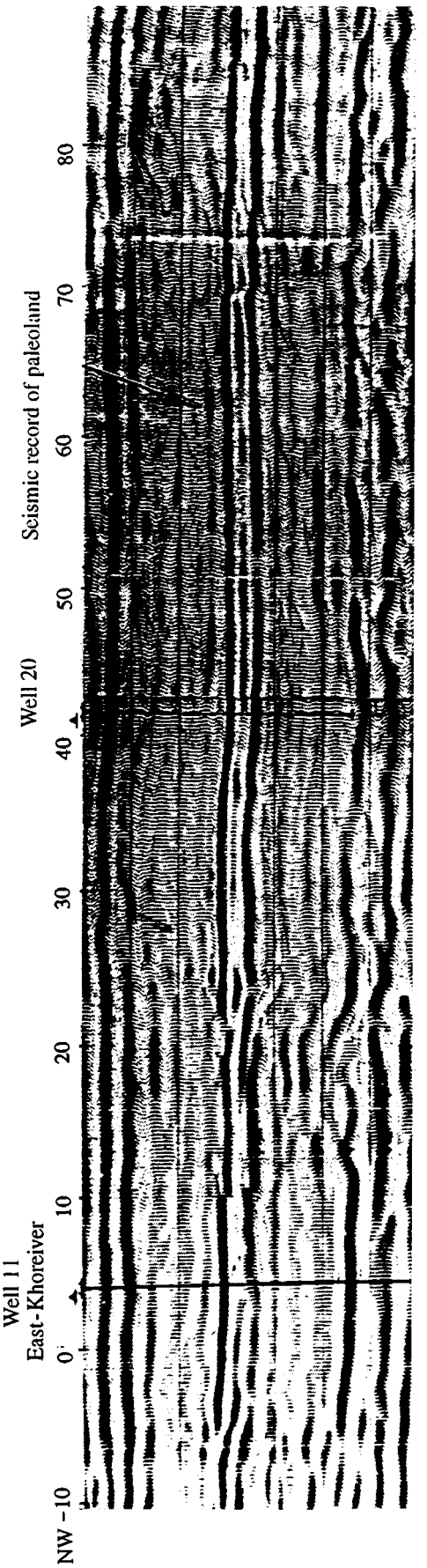

Fig. 4. A fragment of seismic section from the northern Pechora sedimentary basin illustrating the structure of the paleoland deposits.

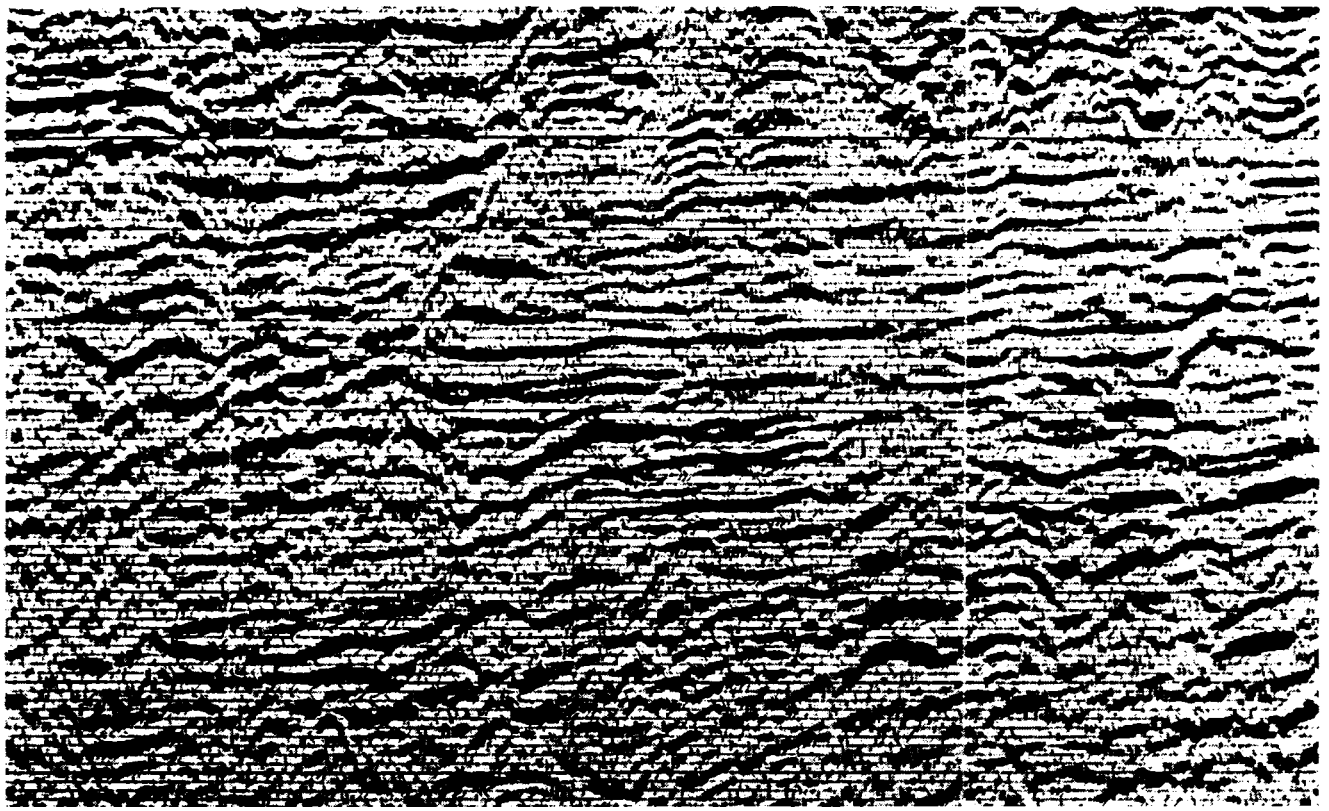

Fig. 5. A fragment of seismic section from northern India illustrating characteristic patterns of molasses.

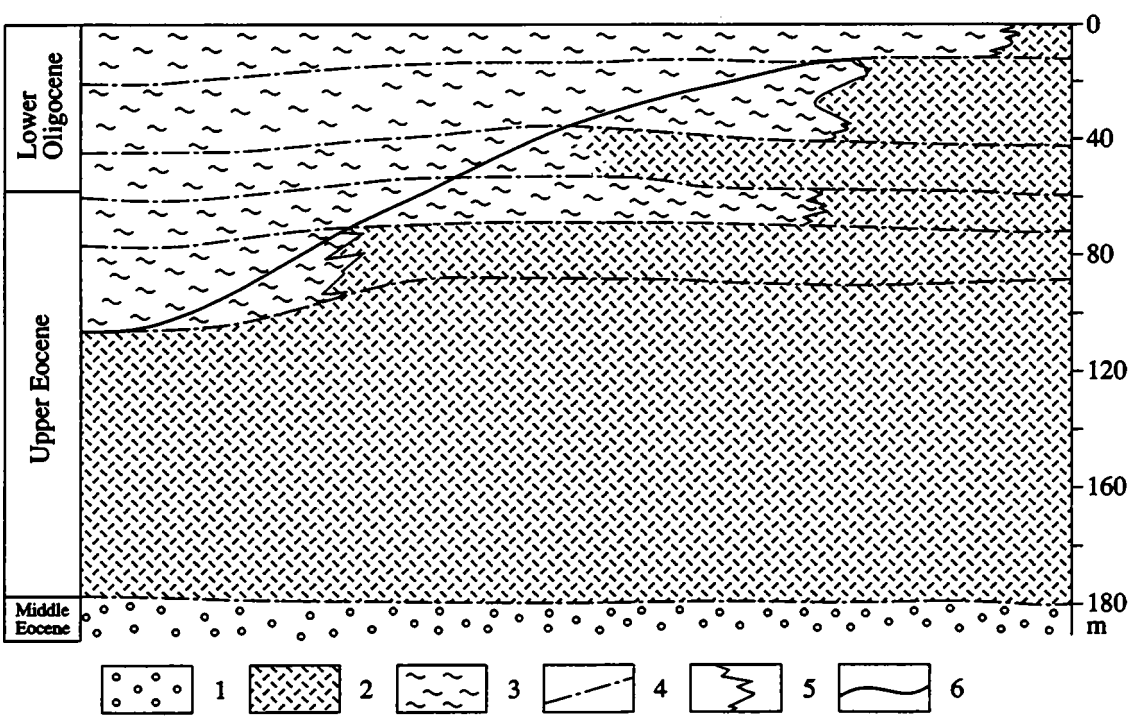

Fig. 6. Correlational scheme of the Saksaul and Chegan Formations in the northern Aral onshore region.
1 – Tasaran Formation; 2 – Saksaul Formation; 3 – Chegan Formation; 4 – isochronous surfaces; 5 – transition between continental deposits of the Saksaul Formation and marine sediments of the Chegan Formation; 6 – boundary between the Saksaul and Chegan formations.

(Fig. 5). Variegated terrestrial deposits are often distinguished by the presence of transparent and semitransparent seismic horizons.

4. MARINE AND CONTINENTAL FACIES AND SEQUENCE CORRELATION

The distinguishing of time seismostratigraphic units is based on the synchronism of seismic reflections. According to their phase correlation, they can be traced across wide areas of a sedimentary basin. However, in the transitional zones between coastal-marine and continental deposits, the amplitude of the reflections, as a rule, sharply decreases and direct phase correlation becomes impossible. Therefore, recognizing pertinent facies in accordance with three groups of seismostratigraphic criteria is the only way to locate and map the transitional zone between marine and terrestrial deposits. Thereafter, one can correlate, without serious mistakes, marine seismostratigraphic units to those of continental sequence.

For example, revealing the structure of a transitional zone between marine and continental deposits of the Upper Eocene–Lower Oligocene age in the northern Aral onshore region based on the described seismostratigraphic parameters. Comparing them with geophysical and biostratigraphic data from the study of boreholes gives new insight into the age interrelations of the Saksaul and Chegan Formations, which have been a matter of contention for decades. The first of the formations (up to 160 m thick) is composed of silty-clay sediments conformably overlying clay deposits of the Tasaran Formation. The upper part of the Saksaul Formation is represented by continental white quartz sands. It is overlain by marine clay of the Chegan Formation (up to 100 m thick). Until the end of the 1960s, it was believed that the Saksaul Formation was of the Late Eocene age, while the Chegan clays were of the Early Oligocene in age (Fig. 6). However, a comprehensive analysis showed that the continental sand units of the Saksaul Formation successively grade into the marine clay of the Chegan Formation from the northeast to the southwest (Fig. 6). Five isochronous levels observed in both marine and continental facies have been revealed using seismic stratigraphy and drilling results. Their junctions represent wedgelike units of facies replacement that mark the location of the land–sea transitional zone in the geological past. Tracing five wedgelike transitional units (up to dozens meters thick) in the Paleogene sequence, we can estimate the stratigraphic ranges of the Saksaul and Chegan formations in the northern and southern parts of the north Aral onshore region. In the north, the boundary between the two formations corresponds to the Upper Eocene–Oligocene boundary, but its level is much lower within the upper Eocene in the south.

Reviewer Yu.B. Gladenkov

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